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# **ON THE SPATIAL AND TEMPORAL DISTRIBUTION OF NEAR SURFACE SOIL MOISTURE ACROSS A LOW-ARCTIC TUNDRA HILLSLOPE**

**David Robert Sayer**

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**Thesis submitted for the degree of Doctor of Philosophy**

**University Of Durham  
Centre for Ecology and Hydrology**

**2007**



**STEPPS**  
Snow in Tundra Environments,  
Patterns, Processes and Scaling



**Centre for  
Ecology & Hydrology**  
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- 5 JUN 2008

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I hereby declare that the contents of this document are entirely my own work unless referenced otherwise.

Signed

A handwritten signature in black ink, appearing to be 'D.R. Sayer', written over a horizontal line.

D.R.Sayer

Date: 10th DECEMBER 2007



## **Abstract**

Recent changes in global climate are having an unprecedented effect on the Arctic environment. This thesis examines the surface hydrology of a heterogeneous low-Arctic Tundra heath hillslope site near Abisko, Northern Sweden (68° 17'N, 18° 51'E) from May 2003 to September 2005. The study centred on two grids; on a hillslope (10800 m<sup>2</sup>) and a flatter site (20000 m<sup>2</sup>) with average slopes of approximately 10° and 7° respectively. Spatial soil moisture measurements, at 5 cm depth and 10 m grid spacing, were carried out in May, June, July, August, September 2003 and May, July, August, September 2004. Time series of soil moisture and temperature (at 5 cm and 10 cm depths) were also recorded at half hourly intervals between June 2003 and September 2005. Spatial surveys of topography, snow depth, soil properties, soil depth and vegetation height were also undertaken producing a high resolution dataset.

Variogram analysis of the spatial soil moisture patterns shows seasonal variation. Ranges of between 50 m and 55 m were obtained, with the variograms having distinct sills and nuggets. The nugget values fall within the range expected due to measurement error. Directional variograms demonstrate that, on the hillslope, topography is an important factor driving the soil moisture distribution during the spring snowmelt period. This is not the case during the drier summer months. Conversely, soil moisture patterns show continuity and connectivity when the soils are wet during the spring snowmelt period.

The topography of the site influences spatial snow distribution, vegetation community structure and soil properties. Topographic index values, derived from digital elevation models, were not correlated well with observed soil moisture values due, in part, to heterogeneous soil properties. This study suggests that the processes driving soil moisture variability for this area switch between topographical controls (when soils are wet) and local controls such as soil properties and evapotranspiration (when soils are dry).

A Monte-Carlo parameter sweep method was employed, using observed soil moisture values, to give effective soil parameters values for use in a land surface model (JULES). The 1D model did not perform well when modelling soil moisture values during the freeze and thaw periods on the hillslope since the effects of lateral water movement could not be accounted for. Comparisons with the flatter site were much more favourable.

It is clear that the hydrology of high latitudes are driven by the same underlying processes as in other parts of the world; the redistribution of water due to topography when soils are wet and the vertical water fluxes due to soil, vegetation and climate when the soils are dry.

### **Acknowledgements**

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## Contents of Thesis

<u>Chapter</u>	<u>Page</u>
1     Introduction and Background.	1
1.1. Climatic Change in the Arctic	2
1.2. The STEPPS Project	2
1.3. Aims of this Thesis	3
1.4. Background Literature	4
1.4.1. Process Basis of Soil Moisture Variability	4
1.4.2. Spatial and Temporal Soil Moisture Variability	6
1.4.3. Soil Moisture and Topography	13
1.4.4. Soil Moisture and Vegetation	17
1.4.5. Modelling Soil Moisture in Land Surface Schemes	19
2     Methodology.	24
2.1. The Study Area	25
2.1.1. The Hydrological Study Area	28
2.2. Soil Moisture Surveys	31
2.2.1. Establishment of Grids	31
2.2.2. The Surface Capacitance Insertion Probe (SCIP)	32
2.2.3. Spatial and Temporal Soil Moisture Measurement	34
2.3. Automatic Data Collection	36
2.4. Fieldsite Surveys	42
3     To investigate the spatial and temporal variability of soil moisture across the Tundra landscape and the processes that control this variability.	46
3.1. Field Methods and Data Collection	47
3.2. Analysis of Data	48
3.2.1. Variogram Production	55
3.2.2. Anisotropy	59
3.2.3. Variogram Modelling	62
3.2.4. Time Series Soil Moisture and Rainfall	70
3.2.5. Soil Freezing and Thawing	75
3.3. Chapter Summary	82

<b>Chapter</b>		<b>Page</b>
4	To assess the role topography plays in determining soil moisture distributions at the hillslope scale.	84
	4.1. Field Methods and Data Collection	85
	4.2. Analysis of Data	86
	4.2.1. Topography and Soil Moisture Distributions	96
	4.2.1.1. The Hillslope Grid	98
	4.2.1.2. The Windfetch Grid	103
	4.2.2. Prediction of Wet Areas	110
	4.3. Chapter Summary	115
5	To determine the relationship between snow depth, soil properties and vegetation height and their relation to soil moisture distributions.	117
	5.1. Field Methods and Data Collection	118
	5.2. Analysis of Data	119
	5.2.1. Snow Observations and Surveys	119
	5.2.2. Soil Surveys	125
	5.2.3. Vegetation Height	128
	5.3. Chapter Summary	130
6	To test the performance of a land surface model for modelling soil moisture in a heterogeneous Tundra landscape.	132
	6.1. Model Description	133
	6.2. Modelling Framework	140
	6.3. Soil Parameter Sensitivity	145
	6.4. Model Performance During Freeze/Thaw Events	155
	6.4.1. The Freeze Period	155
	6.4.2. The Thaw Period	158
	6.5. Evaluation of Model Performance	161
	6.6. Chapter Summary	164

<b>Chapter</b>	<b>Page</b>
7 Discussion.	165
7.1. Spatial and Temporal Soil Moisture Variability	166
7.2. Topographic Influence on Spatial Soil Moisture Patterns	174
7.3. Snow, Vegetation and Soil– Effects on Spatial Soil Moisture Variability	177
7.4. Summary & Conclusions – Soil Moisture Patterns Across a Tundra Landscape	180
7.5. Further Work	183
7.5.1. Soil Depth and Groundwater	183
7.5.2. Spatial Soil Properties	184
7.5.3. Soil Moisture Survey Scales	184
7.5.4. Snowmelt Tracer Experiments	184
7.5.5. Terrain Indices	185
7.5.6. Modelling Process Hydrology for High Latitudes	185

## **Appendices**

### List of References

#### Appendix A: Surface Capacitance Insertion Probe (SCIP) – Calibration, Measurement Error and Comparison with CS616 Water Content Reflectometer

- A.1 Calibration for Soils at the Abisko Fieldsite
- A.2 SCIP Measurement Error and Repeatability
- A.3 Comparison of SCIP and CS616 Water Content Reflectometer Measurements

## **List of Tables and Figures by Chapter**

### **Chapter 1**

**Figure 1.1** The drainage basin hydrological system (from an original diagram by Professor J. Lewin, cited in Ward and Robinson, 2000)

**Figure 1.2** Schematic representation of the MOSES model (showing the 'skin' surface temperature ( $T^*$ ), interactive surface resistance ( $r_s$ ) and soil water phase change)

### **Chapter 2**

**Figure 2.1** Map showing the location of the STEPPS study area.

**Figure 2.2** Photograph showing the STEPPS fieldsite (September 2003).

**Figure 2.3** Photograph showing the features of the Windfetch Grid. View from the southern edge of the grid looking approximately north east (September 2003).

**Figure 2.4** Photograph showing the base of a topographic hollow on the Hillslope Grid. View from the eastern edge looking approximately east-west across the grid (September 2003).

**Figure 2.5** Diagram of the Surface Capacitance Insertion Probe (SCIP) showing the instrument dimensions and characteristics.

**Table 2.1** Dates of the soil moisture surveys across the Hillslope and Windfetch Grids 2003 to 2004.

**Figure 2.6** The Automatic Weather Station (AWS1) on the Windfetch Grid.

**Figure 2.7** The Hillslope Grid soil station (a) and Windfetch soil station (b).

**Figure 2.8** Plot showing the location of measurement sites across the study area.

**Table 2.2** Soil profile sample data (collected by Dr. Baxter and Dr. Wookey, July 2003, unpublished data, pers. comm.).

**Table 2.3** Summary of the field measurements recorded for the Hillslope and Windfetch sites.

### Chapter 3

**Table 3.1** Mean, standard deviation, range and number of data points from the SCIP survey data for the Hillslope and Windfetch Grids.

**Figure 3.1** Monthly average volumetric water content ( $\text{m}^3\text{m}^{-3}$ ) for *Salix* and *Empetrum* Sites on the Hillslope Grid during the Study Period (June 2003 - September 2005).

**Figure 3.2** Frequency histograms of the Hillslope Grid for each spatial soil moisture survey.

**Figure 3.3** Frequency histograms of the Windfetch Grid for each spatial soil moisture survey.

**Figure 3.4** Characteristics of a typical variogram (after Isaaks and Srivastava, 1989).

**Figure 3.5** Omnidirectional variograms for the Hillslope Grid derived from spatial soil moisture survey data.

**Figure 3.6** Omnidirectional variograms for the Windfetch Grid derived from spatial soil moisture survey data.

**Figure 3.7** Directional and omnidirectional variograms of the Hillslope Grid for May and September 2004.

**Figure 3.8** Directional and omnidirectional variograms of the Windfetch Grid for May and September 2004.

**Figure 3.9** Examples of variogram modelling for the Hillslope Grid (spherical model) and Windfetch Grid (exponential model) for May and June 2003.

**Table 3.2** Variogram nugget values for the modeled omnidirectional variograms for the Hillslope (spherical model) and Windfetch (exponential model) sites for the study period.

**Figure 3.10** Grid point naming conventions for the Windfetch Grid and the Hillslope Grid.

**Figure 3.11** Hillslope soil volumetric water content surfaces derived by kriging using spherical variogram models of the spatial soil moisture survey data.

**Figure 3.12 (a and b)** Windfetch soil volumetric water content surfaces derived by kriging using exponential variogram models of the spatial soil moisture survey data.

**Figure 3.13** Recorded rainfall on the Windfetch Grid for the duration of the study period.

**Figure 3.14** Monthly average recorded rainfall (mm) on the Windfetch Grid during the study period (July 2003 – May 2005).

**Figure 3.15** Hillslope Grid volumetric soil water content and recorded rainfall for the *Empetrum* (a) and *Salix* (b) sites during the study period.

**Figure 3.16** Soil volumetric water content at various soil depths and recorded rainfall for the Windfetch Grid.

**Figure 3.17** Hillslope Grid volumetric water content and soil temperature during the study period using CS616 water content reflectometers and T107 soil temperature probes.

**Figure 3.18** Soil volumetric water content and soil temperature at 5 cm and 10 cm depth for the Windfetch Grid.

**Figure 3.19** Soil volumetric water content and soil temperature with depth for the Windfetch Grid.

#### Chapter 4

**Figure 4.1** Location of the Hillslope and Windfetch Grids at the STEPPS fieldsite (displayed in ArcMap).

**Figure 4.2** Trimble Geomatics Office software showing the data points collected on the Hillslope Grid and the format of the point data.

**Figure 4.3** Triangular Irregular Network (TIN) of RTK survey points for the Hillslope Grid displayed in ArcGIS.

**Figure 4.4** Slope maps of the Windfetch and Hillslope Grids.

**Figure 4.5** Surfer 7 wireframe grids of the Hillslope (top) and Windfetch (bottom).

**Figure 4.6** Photographs showing the topography of the Hillslope Grid (taken 23<sup>rd</sup> July 2003).

**Figure 4.7** Photographs showing the topography of the Windfetch Grid (taken 13<sup>th</sup> September 2003).

**Figure 4.8** Photographs of a 'stream' channel to the south (upslope) of the Hillslope Grid (May, July and September 2003).



**Figure 4.9a** Soil volumetric water content (5 cm depth) surfaces mapped onto the Hillslope Grid topography (May and June 2003).

**Figure 4.9b** Soil volumetric water content (5 cm depth) surfaces mapped onto the Hillslope Grid topography (July and September 2003).

**Figure 4.9c** Soil volumetric water content (5 cm depth) surfaces mapped onto the Hillslope Grid topography (May and July 2004).

**Figure 4.9d** Soil volumetric water content (5 cm depth) surfaces mapped onto the Hillslope Grid topography (August and September 2004).

**Figure 4.10a** Soil volumetric water content (5 cm depth) surfaces mapped onto the Windfetch Grid topography (May and June 2003).

**Figure 4.10b** Soil volumetric water content (5 cm depth) surfaces mapped onto the Windfetch Grid topography (July and August 2003).

**Figure 4.10c** Soil volumetric water content (5 cm depth) surface mapped onto the Windfetch Grid topography (September 2003).

**Figure 4.10d** Soil volumetric water content (5 cm depth) surfaces mapped onto the Windfetch Grid topography (May and July 2004).

**Figure 4.10e** Soil volumetric water content (5 cm depth) surfaces mapped onto the Windfetch Grid topography (August and September 2004).

**Figure 4.11** Hillslope (1 m resolution) and Windfetch (1.5 m resolution) Topographic Index values.

**Figure 4.12** Topographic Index (smoothed 5 m values) and soil volumetric water content (SCIP surveys) for the Hillslope Grid during 2003.

**Figure 4.13** Topographic Index (smoothed 5 m values) and soil volumetric water content (SCIP surveys) for the Windfetch Grid during 2003.

## Chapter 5

**Figure 5.1** Photographs of the Hillslope Grid (a) (taken 22<sup>nd</sup> February 2003) and the Windfetch Grid (b) (taken 26<sup>th</sup> February 2003) showing the snow cover at the fieldsite.

**Figure 5.2** Snow survey data for Hillslope Grid (February and May 2003).

**Figure 5.3** Snow depth and land surface elevation along transects of the Hillslope Grid for February and May 2003.

**Figure 5.4** Photographs of a topographic ‘hollow’ on the Hillslope Grid during the spring snowmelt period 2003, (a) 3<sup>rd</sup> May, (b) 15<sup>th</sup> May and (c) 7<sup>th</sup> June.

**Figure 5.5** Averaged soil depths (based on an average of four measurements at each grid) across the Hillslope and Windfetch Grids.

**Figure 5.6a** Hillslope vegetation heights displayed using ArcMap. Grid points are annotated where vegetation height exceeds 19 cm.

**Figure 5.6b** Windfetch vegetation heights displayed using ArcMap. Grid points are annotated where vegetation height exceeds 19 cm.

## Chapter 6

**Table 6.1** Modelling framework describing the objectives of the modelling exercise, description of the model runs and model outputs.

**Figure 6.1** Driving data used in the JULES Land Surface Scheme for the (a) soil parameter, (b) freeze and (c) thaw modelling periods.

**Figure 6.2** Observed soil moisture, temperature and rainfall at the Hillslope *Salix* site for the (a) soil parameter, (b) freeze and (c) thaw modelling periods.

**Table 6.2** Soil parameter values currently used in MOSES/JULES and those suggested for organic soils (Lloyd *et al.*, 1999, cited in Hall, unpublished).

**Table 6.3** Hydraulic conductivity observations (after Letts *et al.*, 2000).

**Table 6.4** Initial parameter sweep upper and lower value limits.

**Figure 6.3** Monte-Carlo parameter sweep output varying  $b$ ,  $\psi_s$ ,  $K_s$ ,  $C_s$  and  $\lambda$ .

**Figure 6.4** Graphs showing the results of Monte-Carlo analysis for varying values of the  $K_s$  (saturated hydraulic conductivity) parameter for three soil moisture dry-down events (slopes) during the summer 2003.

**Figure 6.5** Graphs showing the results of Monte-Carlo analysis for varying values of the  $b$  exponential parameter for three soil moisture dry-down events (slopes) during the summer 2003.

**Figure 6.6** Graphs showing the results of Monte-Carlo analysis for varying values of the  $\psi_s$  (soil water suction) parameter for three soil moisture dry-down events (slopes) during the summer 2003.

**Figure 6.7** Soil temperatures (modelled and observed), air temperature (modelled) and laying snow (modelled) for the freeze modelling period at the *Salix* site.

**Figure 6.8** Photograph showing the *Salix* site on the Hillslope Grid during the freeze modelling period. The snow is seen to persist in the topographic hollow. (Photograph taken on 10<sup>th</sup> November 2003).

**Figure 6.9** Soil temperatures (modelled and observed) and laying snow (modelled) for the thaw modelling period at the *Salix* site.

**Figure 6.10** Modelled soil volumetric water content and laying snow for the thaw modelling period at the *Salix* site.

#### Appendix A

**Figure A1** Volumetric Water Content ( $\text{cm}^3 \text{cm}^{-3}$ ) against Square Root of Relative Permittivity for Abisko Soils using a Surface Capacitance Insertion Probe.

**Figure A2** Volumetric Water Content ( $\text{cm}^3 \text{cm}^{-3}$ ) against Square Root of Relative Permittivity for Abisko Soils using a Surface Capacitance Insertion Probe (removed point at wet end).

**Figure A3** Omnidirectional variogram for the June 2003 SCIP survey showing the estimated spherical model.

**Figure A4** SCIP volumetric water content against CS616 volumetric water content on 29<sup>th</sup> July 2003, 12<sup>th</sup> September 2003, 26<sup>th</sup> May 2004, 29<sup>th</sup> July 2004, 10<sup>th</sup> August 2004 and 4<sup>th</sup> September 2004.

## **Chapter 1**

### **On the Spatial and Temporal Distribution of Near Surface Soil Moisture Across a Low-Arctic Tundra Hillslope**

#### **Introduction and Background**



This study is concerned with the spatial and temporal distribution of near surface soil moisture across a low-Arctic Tundra hillslope. Of particular interest are the periods during the seasonal freezing and thawing of the soils and the role of topography on the soil moisture variability as this will provide an insight into the factors controlling soil water redistribution across the landscape.

### 1.1 Climatic Change in the Arctic

The tendency for Arctic Tundra to act as a long-term carbon sink and the presence of high albedo surfaces in such areas make them extremely valuable moderators of the global energy budget. Tundra environments are highly vulnerable to the effects of atmospheric warming; changing spatial extents of high albedo surfaces could lead to a switch from carbon sink to a source of atmospheric carbon generating a positive feedback which could exacerbate global warming (Intergovernmental Panel on Climate Change (IPCC), 2001). The hydrology of heterogeneous Tundra landscapes is an important aspect of such ecosystems, having a large influence on soil biogeochemical processes and vegetation community structure (Rodriguez-Iturbe, 2000).

### 1.2 The STEPPS Project

The Snow in Tundra Environments: Patterns, Processes and Scaling (STEPPS) project was established in 2002 in order to investigate the impacts of global warming on Tundra ecosystems. Funded by the Natural Environment Research Council (NERC), the project is an interdisciplinary venture with contributors from the Centre for Ecology and Hydrology, Wallingford; the School of Biological Sciences, University of Durham; the Sheffield Centre for Arctic Ecology, University of Sheffield and the Department of Physical Geography, University of Uppsala, Sweden.

The STEPPS project aims to provide an understanding of the biogeochemical processes operating in Tundra regions in order to investigate the impacts of global warming on these ecosystems. The project has four main hypotheses, one of which states:

*That a change in the depth/duration of snow-lie will, as a result of direct effects upon soil thermal and hydrological regimes, alter the thermal balance of the soil and hence result indirectly in altered seasonal dynamics of soil biogeochemistry.*

In order to investigate the effect that the depth and duration of snow-lie has on the soil thermal and hydrological regimes, this PhD thesis aims to understand and quantify spatial and temporal patterns of soil moisture across a low-Arctic Tundra landscape and the controlling factors (such as topography, soil depth, snow cover and vegetation type) that determine these distributions.

### 1.3 Aims of this Thesis

Two main research aims have been identified in order to structure the hydrological investigations. These are:

1. To investigate the spatial and temporal variability of soil moisture across the Tundra landscape and the processes that control this variability. In particular:
  - a. To assess the role topography plays in determining soil moisture distributions at the hillslope scale.
  - b. To determine the relationship between snow depth, soil properties and vegetation height and their relation to soil moisture distributions.
2. To test the performance of a land surface model for modelling soil moisture in a heterogeneous Tundra landscape.

These research areas have been explored by extensive field measurement campaigns during which data were collected on the spatial and temporal distributions of soil moisture, the influence of a spatially distributed snowpack on soil moisture patterns and the processes at work during freeze/thaw cycles. Also, the link between the spatial distribution of vegetation height, soil depth, soil moisture and topography were explored.

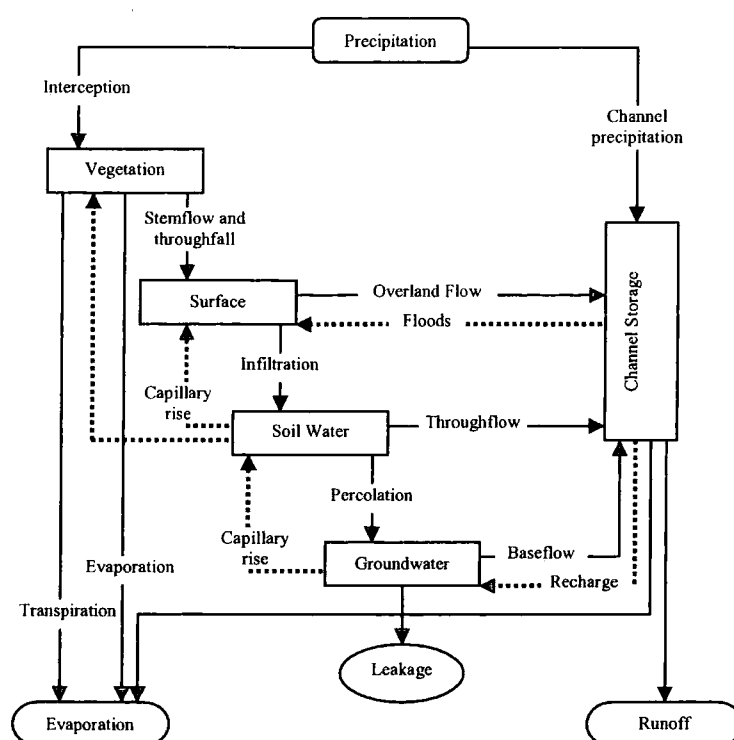
Computer modelling techniques using the Meteorological Office Surface Exchange Scheme (MOSES) and Joint UK Land Environment Simulator (JULES) Land Surface Schemes were used to determine effective soil parameter values for the Tundra soils and evaluate model performance during freeze/thaw cycles across the Tundra heath hillslope.

#### 1.4 Background Literature

There are many complex issues that need to be addressed when investigating the hydrological processes of Arctic regions. In this section, a brief introduction into the process basis of soil moisture variability on low-Arctic hillslopes will be presented and views on the state of high latitude hydrological research will be addressed. A number of field studies will then be outlined which give an insight into the current research on the spatial variability of soil moisture patterns and the influence of topography on these patterns. Current issues in the field of hydrological modelling will then be outlined.

##### 1.4.1 Process Basis of Soil Moisture Variability

Soil water is a very important aspect of the hydrological cycle which is concerned with the water contained in the soil profile and the subsurface water in the layers between the soil profile and the water table. Figure 1.1 shows a representation of the hydrological processes at the catchment scale.



**Figure 1.1 The drainage basin hydrological system (from an original diagram by Professor J. Lewin, cited in Ward and Robinson, 2000)**

When a precipitation event occurs, the water that reaches the soil surface is absorbed unless the soil is saturated, in which case the water will flow over the surface (termed overland flow). Ward and Robinson (2000) describe the classification of subsurface water into four zones. The water table coincides with the upper surface of the zone of saturation and above this is the capillary fringe where all the soil pore spaces contain water. In the intermediate zone (above the capillary fringe) the predominant direction of water movement is downwards from the soil zone where the precipitation enters the soil.

The soil itself comprises a number of different horizons which each have different physical properties. Water is stored in the pore spaces between soil particles due primarily to the forces of capillary action, adsorption and osmosis. The distribution of these pore spaces greatly influences soil water movement and retention within the soil.



The input of water through snowmelt into the hydrological system is particularly important in higher latitude studies. The spatial distribution of snowfall, amount of cover and melting processes are particularly important to hydrologists in these areas. When considering the hydrology of regions in high latitudes, the presence of frozen soils should also be considered. When snow melts in these regions, the infiltration of water into frozen soils, or the lateral redistribution over the frozen soil surface, may contribute greatly to the hydrological regime since the melt period is the most significant annual influx of water. At the hillslope scale, Dingman (1994) suggests that there are three main methods of snowmelt runoff generation. These are the infiltration of meltwater into the soil beneath the snowpack, the downslope redistribution of meltwater at the base of the snowpack due to soil saturation or frozen ground or a combination of the two.

#### 1.4.2 Spatial and Temporal Soil Moisture Variability

The field measurement of spatial and temporal soil moisture patterns is essential in order to quantify hydrological processes and to aid process understanding.

There have been many studies into the spatial and temporal patterns of soil moisture at a number of sites worldwide. A summary of a number of these studies can be found in Western *et al.* (1998a). The summary provides details of the location, area, depth, number of sample points, date and statistics such as variance, mean and range. Some of the studies covered in this summary are outlined below.

The spatial distribution of soil moisture can be described in many ways. Western *et al.* (2003) outline the processes and statistical techniques for the representation of soil moisture patterns. The way in which spatial patterns may be random (variability that has predictable statistical properties but can not be predicted in detail) or organised (variability that has regularity) are discussed. Western *et al.* (2003) explain that organised variability may display continuity (that can be captured statistically), connectivity (connected bands such as drainage lines) or convergence (such as branching drainage lines). These organised

features may be displayed under a number of circumstances and they found that the degree of organisation depends ultimately on the soil moisture state.

The spatial organisation of soil moisture fields is also described by Rodriguez-Iturbe and Rinaldo (1997). In this case, the emphasis is placed on investigating the links between the properties of soil moisture fields (such as measured soil moisture, soil texture, land use and soil-atmosphere interaction) when observed at varying degrees of spatial resolution. Remotely sensed soil moisture (and soil texture) data were used (from the Washita River watershed in Oklahoma, U.S.A) to give soil moisture fields of  $200 \times 200 \text{ m}^2$ ,  $400 \times 400 \text{ m}^2$  and  $600 \times 600 \text{ m}^2$  resolution. Gravimetric soil moisture and soil porosity data were also collected at sixteen sites ( $30 \times 30 \text{ m}$ ). Variance and correlation analysis of the soil moisture fields suggest that their structure is typical of scaling processes which do not change the statistics of their fluctuations at different levels of resolution (Rodriguez-Iturbe and Rinaldo, 1997). The authors suggest that soil-atmosphere interaction, rainfall or surface runoff would not be related to the spatial scaling of the soil moisture fields because of the size of the study area and timescales used in the study. However, the porosity of the soils at the scales discussed did show a scaling spatial field and it is suggested that the spatial scaling structure of soil moisture fields may arise from existing scaling in soil properties.

One approach to understanding the hydrological processes operating at a site is to undertake water balance studies. To provide a quantitative assessment of the movement of water through the system, these studies are based on the law of conservation of mass that states that the change in storage of water in a system will be equal to the input minus the output (Ward and Robinson, 2000). Rovaneck *et al.* (1996) describe one such investigation based on field measurements made on the Alaskan Arctic Coastal Plain near Prudhoe Bay. Here, snowmelt was the main driving factor for the hydrological process and the study shows that the persistent presence of ponds and wetlands was dependent on snowmelt from the uplands. The wetlands lost water over the summer due to evapotranspiration and were filled to capacity during the spring snowmelt. Surface flow was only observed at their site following snowmelt and, in this case, the presence of the

snowpack made the water balance calculation impractical in their study area during snowmelt (Rovansek *et al.*, 1996).

Carey and Woo (2001) carried out water balance studies for a sub-Arctic hillslope in the Wolf Creek Basin in the Yukon. At this site, snow accounted for half the annual water input into the system. Carey and Woo studied the water balance for hillslopes with different aspects and concluded that the snowmelt timing and rate varied by up to two months on hillslopes with similar climate. Slopes with organic soils and ice at the base produced runoff during snowmelt periods and the evapotranspiration was the largest water loss from the hillslopes. The rainfall-runoff studies carried out by Spence and Woo (2002) in the Northwest Territories of Canada suggest microtopography, soil characteristics and climate all influence rainfall-runoff relationships of the bedrock slopes studied in this area. They suggest that the depressions in the bedrock provided storage and lengthened residence time at the expense of runoff.

These water balance studies provide useful information about the hydrological status of these areas. However, it is the field-based determination of the spatial distribution of soil moisture and the variability of moisture patterns across the landscape that is most relevant to this thesis since these studies can give detailed information on the movement of water at the hillslope scale.

In 1969, Hills and Reynolds described hydrological investigations carried out in a small experimental catchment near Chew Stoke, south of Bristol in the UK. This study was undertaken in order to evaluate infiltration rates and sampling strategies (specifically to determine the number of samples needed to represent the mean soil moisture accurately). Hills and Reynolds suggest that, based on their field experiments, a large number of samples (80 in this study) should be taken to give an estimate with 95% probability. It is also suggested that the soil moisture variability could be assessed from the surface vegetation and soil structures present and that the slope angle and degree of variability are related.

Temporal variations in soil moisture were investigated at two sites (one 'wet' and one 'dry') in the Wye catchment (UK) by Roberts and Crane (1997). The wet site was located in a flat (mire) area dominated by heath land overlying peat whilst the dry site was on a well drained slope with brown earth soils. The study showed that little temporal variation in soil moisture occurred at the wet (flat) site. The dry (slope) site showed major variations in soil moisture content over time with significant moisture deficits during dry periods. This was due to the soil moisture content on the slope responding rapidly to storm events but during dry periods the soil moisture content showed little variation. The authors suggest that, during storm events, soil water is redistributed laterally downslope as sub-surface flow into the flatter (mire) area which expands upslope. Although this study was carried out in a temperate region, the processes described in this study at the hillslope scale could be similar to those operating in other climates such as the low-Arctic.

In the United States, a number of studies have described the variability of soil moisture distributions. Bell *et al.* (1980) describe the results of soil moisture data collection at a number of fieldsites in Arizona, Kansas and South Dakota. At these sites, gravimetric soil moisture surveys were collected and analysed to determine the relationship between soil moisture variability and mean soil moisture. The study showed that the variability-mean data pairs were clustered about a mean soil moisture (for a given site) and recommended that more surveys carried out over time would provide a better estimate for this relationship.

The measurement of soil moisture is no longer restricted by the ability to undertake extensive manual surveys in the field. The use of remote sensing, by sensors on aircraft or satellites is becoming increasingly widespread. The Washita '92 experiment was carried out to test the ability of airborne passive microwave radiometers to measure soil moisture over the Little Washita River Watershed, Oklahoma. The variability of the soil moisture was evaluated by Cosh *et al.* (2004) using moment and regression analysis. They concluded that there was no skewness in soil moisture distribution and that scaled variability over the drying period remained constant. The regression analysis showed that vegetation density,

soil texture and time since precipitation accounted for the majority of soil moisture variability across the watershed.

Soil moisture data collected in Virginia allowed Albertson and Montaldo (2003) to develop a conservation equation for the spatial variance of root zone soil moisture. Although primarily a modelling exercise using remotely sensed data, this study evaluated the soil moisture data and concluded that soil texture distributions imprints variability on the soil moisture at their site.

In Idaho, Grant *et al.* (2004) investigated the spatial and temporal variation of soil moisture for a mountain catchment dominated by snow cover. Manual measurements were carried out using neutron probes across the catchment over two years and the study used mean and standard deviation statistics to describe the spatial variability of the soil moisture. Grant *et al.* suggest that, at their site, the spatial variability of soil moisture was greater than that of other studies due to the heterogeneous conditions and that their study highlighted the difficulty of characterising soil moisture patterns due to this heterogeneity.

The temporal variability of near-surface soil moisture in Québec, Canada was investigated by Parent *et al.* (2006). Here, fine resolution time series data were analysed using a wavelet method. Similarities between moisture variability along the measured transect was attributed to soil homogeneity. The wavelet analysis suggested that, at the 1 to 48 hour scale, soil moisture variability was linked to rainfall events and their intensity and duration. However, the majority of soil moisture variation was accounted for at the 48 hour to 2 week scale where it was the periodicity of rainfall events and duration of dry spells that affected the soil moisture state. This was attributed to the energy transfer from the rainfall to the soil moisture which is dependant on the soil properties. This study suggests that soil moisture variation at the two week temporal scale is linked to the properties of the soil (such as texture and porosity) which effect the re-wetting of the soil after drying.

However, a study carried out by Loague (1992) at the R-5 catchment in Oklahoma compared two datasets; one carried out over 6 years and the other over 6 days. The study

suggests that in the 6 year dataset, soil moisture content varied greatly both spatially and temporally and was not related to the soil type in this catchment. Loague concludes that no spatial structure was observed in the 6 day dataset.

The influences of topography and soil properties on spatial variability of soil moisture are highlighted by Famiglietti *et al.* (1998). Their study, in Austin, Texas, shows that during storm events, the redistribution of soil water is mainly due to topographic and soil properties and that in wet conditions the soil moisture variability is due to the properties of the soil (such as porosity and hydraulic conductivity). Their study suggests that, in dry conditions, the relative elevation, aspect and soil clay content exert a greater influence on the soil moisture distribution. Petrone *et al.* (2004) also concluded, in their study of soil moisture variability in Québec, that the pattern of soil moisture during wetting and drying cycles differs. This study also highlights the link between soil moisture distributions and vegetation cover.

In 1995, a field experiment was initiated in the 10.5 ha Tarrawarra catchment in southern Victoria, Australia. The aim of this experiment was to collect a comprehensive dataset of spatial and temporal soil moisture patterns using a suite of measurements including Time-Domain Reflectometry (TDR), Neutron Moisture Meters (NMM) and remote sensing (Western *et al.*, 1999a).

The statistical exploration of soil moisture distributions for the Tarrawarra catchment has been widely documented (Grayson *et al.*, 1997; Western *et al.*, 1998a; 1998b; 1999a; 2004; Wilson *et al.*, 2004). In many of these, the spatial correlation between measurements as a function of separation distance was described using the variogram geostatistical method. The use of indicator variograms, where thresholds are used to assign indicator values, were evaluated for this catchment but the results showed that this method was not appropriate if the connectivity of hydrological features were to be identified (Western *et al.*, 1998b). The results of the standard variogram analysis, however, were better. These show that high sills and low correlation lengths were observed during wet periods and low sills and high correlation lengths during dry periods (Western *et al.*, 1999a). Two preferred hydrological

states were identified based on the Tarrawarra observations. The first, the wet state, was determined by the lateral flow of water through surface and subsurface flow paths causing the wet areas to occur along drainage lines due to the catchment terrain. In the second, the dry state, the spatial patterns of soil moisture were influenced by vertical fluxes, soil properties and only local topographic effects (Grayson *et al.*, 1997).

Anctil *et al.* (2002), in their study of spatial soil moisture in Québec, Canada, suggest that the results of their variogram analysis compare favourably with those outlined by Western *et al.* (1998a) and that the higher sills observed in their study were due to the presence of organic soils.

The Tarrawarra data were also used, along with data collected from the Hahurangi River catchment on the North Island, New Zealand to compare the spatial soil moisture processes to topographically driven subsurface lateral flow (Western *et al.*, 2004). In this case, the spatial variance of soil moisture was found to be related to the mean soil moisture status and that similarities between the structure of the soil moisture fields and topographic indices representing lateral flows exist. The results of this study suggest that similar conditions within a catchment (such as spatial mean soil moisture and time of year) would produce similar patterns (Western *et al.*, 2004). However, Tromp van Meerveld and McDonnell (2005) do not agree with this view and presented data from their studies at the Panola Mountain Research Watershed near Atlanta Georgia, U.S.A which appear to show that subsurface saturation is related not to soil moisture patterns, as suggested by Western *et al.* (2004), but to soil depth and bedrock micro-topography. Tromp van Meerveld and McDonnell suggest that it is the connection of areas of transient saturation, controlled by the bedrock topography, which controls the generation of lateral subsurface flow. In reply, Western *et al.* (2005) emphasise that the topography is not the only controlling factor of lateral flow and soil moisture patterns. While agreeing with Tromp van Meerveld and McDonnell that in some cases where drainage is rapid, the soil moisture may not be a good indicator of lateral subsurface flow, Western *et al.* reiterate that in areas that saturate over long periods, the soil moisture pattern may be a good indicator of lateral flow. Also, the

importance of other spatially (and temporally) varying properties and processes such as soil properties and evaporative forcing are emphasised by Western *et al.* (2005).

The view that the spatial and temporal distribution of soil moisture is dependent on factors such as the distribution of soils and vegetation, and not topography alone is put forward by Wilson *et al.* (2004). This study, again using data from the Tarrawarra and New Zealand sites, examined the spatial and temporal behaviour of soil moisture processes and concluded that seasonality was the dominant source of variability at these sites and the spatial distribution of soils and vegetation were of equal importance to that of topography.

It has become clear, from a number of these studies, that the links between spatial soil moisture distributions and factors such as mean soil moisture, soil properties, vegetation cover and topography are evident. However, Famiglietti *et al.* (1998) call into question the strength of our understanding of these links, highlighting a number of articles that show no such correlation between spatial soil moisture variability and these factors.

This brief overview of spatial soil moisture field studies has shown that the collection of soil moisture data in the field is of utmost importance in order to gain an understanding of the hydrological processes operating in the real world. With the emphasis in recent years being placed more on modelling hydrological systems than data collection, these studies show that, in order to appreciate the processes involved, the observation, collection and analysis of field data is still relevant and essential.

#### 1.4.3 Soil Moisture and Topography

In recent years it has become increasingly popular to investigate the controls on hydrological systems due to topography. In 2001, Grayson and Western commented on this and questioned the validity of some studies that exploit the ease with which computer simulations can be used to draw conclusions about soil moisture distributions and topography. They argue that, although there are situations where topography does influence factors such as soil moisture distribution, water table depth and areas producing runoff,



these situations are, in fact, not commonplace. It is suggested that it may be more realistic to separate the landscape into two distinct regimes; gully areas that accumulate water which affect the soil moisture state and secondly areas on hillslopes where, if topographic influence exists, it may be due to aspect or variability of soil properties. Grayson and Western make the point that the simulation of topographic influences on soil moisture should only be used because they represent real patterns and processes.

Many studies have highlighted the influence of topography on runoff and throughflow (Anderson and Burt, 1978; Uchida *et al.*, 2006; Freer *et al.*, 1997; 2002). However, it is the role that topography plays in the spatial distribution of soil moisture that is most relevant to this thesis and will be explored here.

Engstrom *et al.* (2005) present a novel system to explore the relationship between micro-topography and near-surface soil moisture distribution at their site in Barrow, Alaska. Here, the land surface was split into 'microsites' based on their curvature and then the soil moisture means and standard deviations calculated for each. The study showed that the micro-topography did control the distribution of near-surface soil moisture in troughs and over small scales (up to 10 metres) and relative topographic position accounted for the variation at larger scales (10-100 metres).

Crave and Gascuel-Oudou (1997) used gravimetric soil moisture sampling in the Coët-Dan catchment in Brittany, France to measure soil moisture distributions along ten transects parallel to slopes. Their analysis suggests that the topography does play a role in moisture distributions but of equal importance was the distribution of soil properties. The difference between upslope and downslope soils caused Crave and Gascuel-Oudou to conclude that downslope topographic conditions may be the controlling factor of surface soil moisture distribution within the catchment. This was due to the determination of two contrasting domains; an upper domain characterised by almost constant soil moisture which varied slowly and a lower domain with higher moisture status and variability.

The evaluation of the influence of topography on spatial soil moisture distributions is often achieved by the use of distributed topographic models. These models use terrain indices to predict the occurrence of wet areas based on topographic features such as slope, curvature and upslope contributing area. Evaluations of some of these methods can be found in Barling *et al.*, (1994), Wigmosta and Lettenmaier (1999), Svetlitchnyi *et al.* (2003), Güntner *et al.* (2004) and Sorensen *et al.* (2006).

Burt and Butcher (1985) examined two terrain indices, the first based on the slope and contributing upslope area proposed by Beven and Kirkby (1979, cited in Burt and Butcher, 1985) and the second based on curvature derived from gridded altitude data (Evans, 1980 cited in Burt and Butcher, 1985). Soil moisture data from the Slapton Wood catchment in south Devon were used to evaluate the two indices using standard statistical techniques (correlation coefficients). Both indices did not perform well in this study and the best results were obtained by using a combination of the two. However, the combined index only predicted soil moisture distributions accurately when the soils were wet; the index performed poorly when the soils on the slope were dry.

One terrain based hydrological model has become used extensively to evaluate and predict topographic influences on hydrological systems. TOPMODEL (a TOPographically based hydrological MODEL) uses the topographic index to determine areas of hydrological similarity. First suggested by Kirkby and Weyman (1974, cited in Beven, 1997), the topographic index,  $k = \alpha / \tan \beta$  (where  $\alpha$  is the upslope area draining through a point and  $\tan \beta$  is the local slope angle), is an intrinsically simple concept which has made it very popular. High values of topographic index indicate areas that will tend to saturate first. A detailed description of TOPMODEL and other terrain indices can be found in Beven (2001), along with examples of studies using terrain indices.

Beven (1997) provided a critique of TOPMODEL, outlining the basic assumptions of the model itself, model parameters, model validation and future applications. Beven makes clear in this critique that the model does have limitations, namely that the use of the topographic index simplifies the catchment dynamics and the model assumptions may be

restrictive. He warns that the model should be used with care and that the model may not be suitable for use everywhere.

This aside, numerous studies have used the TOPMODEL approach, or derivatives thereof, to investigate the topographic influence on factors such as soil moisture distributions, water table variations and subsurface flows (Moore and Thompson, 1996; Seibert *et al.*, 1997; Stieglitz *et al.*, 1999; Walter *et al.*, 2002; Kavetski *et al.*, 2003; Hjerdt *et al.*, 2004; Bormann, 2006 and Cai and Wang, 2006).

Blazkova *et al.* (2002) question the validity of TOPMODEL predictions of distributed water table for the Uhlirská catchment in the Czech Republic. The predictions were evaluated using global catchment parameters and allowing parameter adjustments. The results indicated that water table measurements showed consistent functional differences from the TOPMODEL predictions, mainly due to the simplified dynamics of the saturated and unsaturated zone in the model.

The performance of TOPMODEL in predicting spatial soil moisture fields has been questionable. Some studies have indicated that the model performs best in areas of high soil moisture and not in drier areas. Blyth *et al.* (2004) show that the application of TOPMODEL principles to two catchments, the Wye and the Pang, produced differing results. The topographic index predicted the soil moisture well for the Wye, where peat and gley soils are present, but not for the Pang which has quick draining soils.

However, Nyberg (1996) presented data from the covered catchment at Gårdsjön, Sweden, which suggested that the topographic index indicated that the soil moisture was strongly correlated with topography, despite a wide range of water content (5 – 60%) and sandy-silty soils.

Western *et al.* (1999b) discuss the relationship of observed spatial organisation of soil moisture to terrain indices for the Tarrawarra catchment. They conclude that the wetness

indices should take into consideration the processes occurring at a given catchment under given climate conditions.

This is taken into account by Wilson *et al.* (2005) who propose a method that predicts spatial patterns of soil moisture based on the changing significance of spatial attributes to average wetness status. Three applications were tested; one with only topographic data, one with topographic data plus residuals representing additional spatial data and one with the previous attributes but with an added uncertainty term. The results of their analysis showed that the model performed best with added spatial data and the uncertainty term suggesting that any additional information about a study area that can be added to a terrain index will improve the performance of that index.

Ultimately, the use of topographic indices for the prediction of spatial hydrological patterns is useful when the assumptions that they are based on apply to the study in question and the user should be aware of the limitations of the model. Model development, calibration and testing using observed data will make the output more reliable and applicable to real world situations. Grayson *et al.* (2002) discuss these issues and offer methods aimed at improving the performance of such models based on observed data.

#### 1.4.4 Soil Moisture and Vegetation

Low-Arctic Tundra typically occupies a region north of the boreal forest treeline and the discontinuous patch Tundra of the high-Arctic. The vegetation structure in these regions is spatially heterogeneous and is governed by a number of factors including climate, soil properties and geomorphology. The links between available soil moisture and ecologic patterns have been discussed by a number of researchers (Löffler, 2005, Pheonix and Lee, 2004, Rodriguez-Iturbe *et al.*, 1999, Rovanseck *et al.*, 1996). However, there are two main areas investigated in the literature that are particularly relevant to this study.

The first is the interdependence of climate-soil-vegetation dynamics and the importance of soil moisture on these interactions. Rodriguez-Iturbe (2000) describes the role of soil

moisture in terms of the infiltration, evapotranspiration and leakage (or deep infiltration) and highlights the dependence of the soil moisture state on all of these factors. In particular, Rodriguez-Iturbe (2000) explains that the dependence of evapotranspiration on relative soil moisture depends on the type of vegetation at a particular site. Also, the stress that vegetation undergoes due to water deficits has a direct impact on spatial interactions between vegetation at neighbouring sites. The importance of water use by differing plant communities is highlighted and Rodriguez-Iturbe (2000) suggests that spatial and temporal patterns of soil moisture are both cause and consequence of regional vegetation patterns. This has particular relevance to this study since changes in soil moisture patterns across the heterogeneous Tundra landscape of the low-Arctic could potentially affect the spatial patterns of Tundra plant communities and vice-versa.

The second is the influence of spatially distributed snow on vegetation patterns in the Tundra regions. The hydrology of the low-Arctic, at the hillslope scale, is dominated by the seasonal presence of spatially distributed snow cover and seasonally frozen soils which, during the spring thaw period, are the main source of soil water input into the system. The spatial distribution of snow cover is mainly due to the effect of wind redistribution across the landscape and the accumulation of snow in topographic hollows and areas of tall vegetation. Sturm *et al.* (2005) highlight the effect of tall shrubs on the spatial distribution of snow in Tundra environments. Their research suggests that the snow in the shrubs was considerably deeper (17% - 48%) than that observed on nearby shrub-free Tundra. This effect has a number of consequences on the soil-vegetation dynamic. The snowpack tends to insulate the underlying soil, keeping soil temperatures higher than areas that are snow free. This has implications on winter microbial activity, increasing this process and its duration. Sturm *et al.* (2005) suggest that Tundra regions could be converted to shrubland due to rising air temperatures and increased winter microbial activity. A complex feedback loop is described where soil temperatures are higher in shrubland, resulting in higher microbial activity, which causes more net nitrogen mineralization during the winter and higher shrub leaf nitrogen in the summer. This may increase the decomposability of the litter, favouring shrub growth, which traps more snow leading to higher soil temperatures (Sturm *et al.*, 2005). Any increase in taller shrubs in this region would also have an effect

on the carbon budget due to changes in net carbon exchange and carbon storage. The increase in distributed snow cover due to the abundance of taller vegetation would also have a consequence on soil moisture patterns during the spring thaw period. This, in turn, could influence the spatial patterns of plant communities.

Hence, the determination of soil moisture variability across the low-Arctic Tundra landscape is an important factor when investigating the ecology of these regions and any potential changes in community structure associated with climate warming.

#### 1.4.5 Modelling Soil Moisture in Land Surface Schemes

The use of land surface schemes to represent land-atmosphere interactions within climate models is widespread. A number of different schemes have been developed, some designed for use within global models and some stand-alone (Abbott *et al.*, 1986; Ostendorf and Reynolds, 1993; Ducharne *et al.*, 1998; Kuo *et al.*, 1999; Bronstert and Bárdossy, 1999; Naden *et al.*, 2000; Martinez and Duchon, 2001; Herbst and Diekkrüger, 2003; Pathmathevan *et al.*, 2003, Chen and Hu, 2004 and Isham *et al.*, 2005).

Polcher *et al.* (1996) carried out a comparison of three land surface schemes designed to be used in General Circulation Models (GCM). The schemes, used in the GCM's of the Max-Planck Institute für Meteorologie (MPI), the Hadley Centre-Meteorological Office (UKMO) and the Laboratoire de Météorologie de Dynamique (LMD), were evaluated to analyse their performance to predict the seasonal hydrological budget for two cases; a humid climate with vegetation cover and a dry climate with dry soil. Polcher *et al.* concluded that simulated evaporation, runoff and soil moisture were different for each model. The study concluded that, by changing land surface parameters, the differences between the schemes could be reduced.

The intercomparison of land surface schemes is widely used to assess model performance. The Project for Intercomparison of Land-surface Parameterisation Schemes (PILPS) was a multi-phase exercise that compared the performance of sixteen land surface schemes.

Pitman *et al.* (1999) describe the results of one phase of modelling using data representative of a tropical forest and mid-latitude grassland. They concluded that the models did not agree well and called into question the reliability of land surface schemes.

However, many of the schemes available are being constantly evaluated to improve their performance. Intercomparison studies such as those outlined above go some way in determining the shortcomings of the schemes. The calibration and parameterisation of hydrological processes within many schemes is ongoing and forms a crucial part of the model development (Gedney *et al.*, 2000; Crossley *et al.*, 2000; Walker and Houser, 2001; Nykanen and Foufoula-Georgiou, 2001; Mertens *et al.*, 2004 and McCabe *et al.*, 2005).

Seibert and McDonnell (2002) suggest that the calibration of models should involve input from experimentalists so that 'soft data' (i.e. qualitative data on the observed processes at particular sites) can be used to improve model performance. These 'soft data' may be in the form of groundwater level data or spatial extents of hydrological zones and could be incorporated into conceptual models to improve process understanding and ultimately model performance.

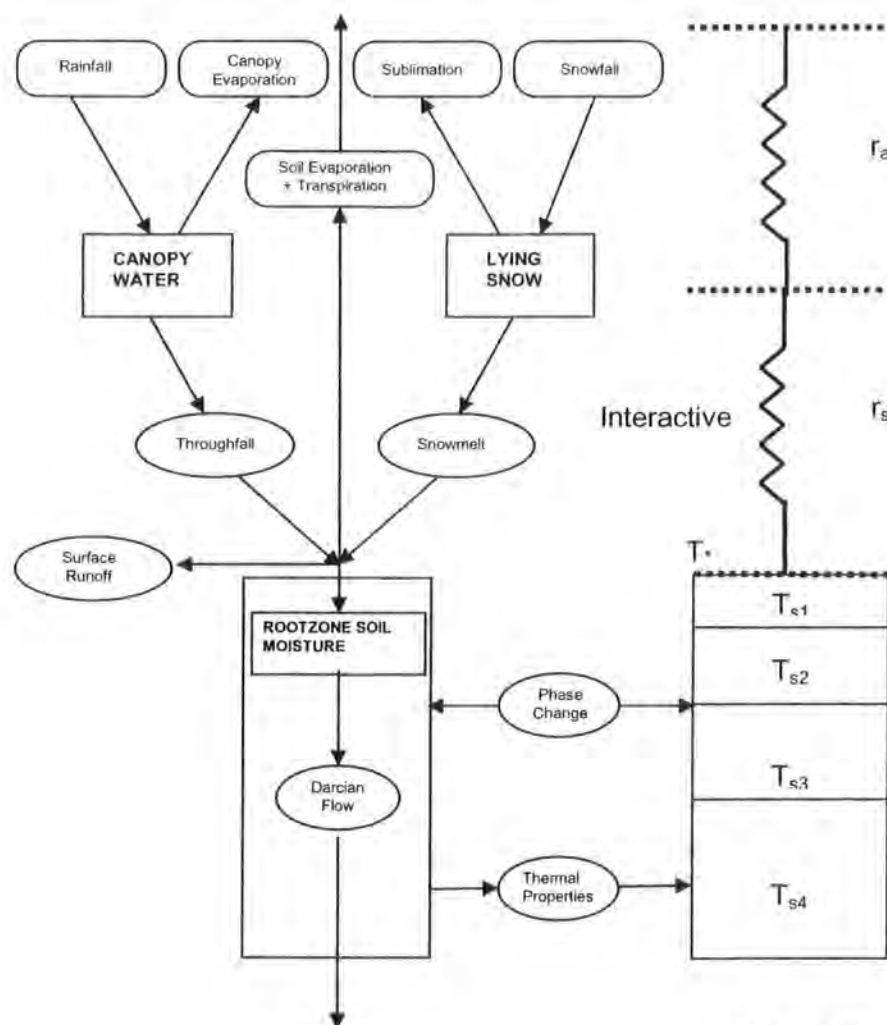
It should be noted that the performance of land surface schemes and climate models depends on the climate being modelled and it is very difficult to compare model performance across different environments.

One of the aims of this thesis is to evaluate the performance of a land surface scheme for high latitude studies. The scheme chosen is the Meteorological Office Surface Exchange Scheme (MOSES), first introduced by Cox *et al.* (1999). MOSES was developed for the UK Meteorological Office (UKMO) climate model and provides a one dimensional description of the exchanges of water, energy and carbon dioxide at the land surface. A schematic diagram of the MOSES structure can be seen in Figure 1.2.

The MOSES scheme has a 'tile' structure which allows different surface types to be represented in each grid square and the surface fluxes are then averaged over the grid square depending on the proportion of each surface type present (Lloyd *et al.*, 1999). The soil is treated as a one-dimensional system and does not account for any lateral flow; only vertical processes are represented in the scheme. The soil component of MOSES consists of a four layered soil model in which the movement of soil water and heat between the layers is calculated. The four soil layers extend down to 3 metres depth, with boundaries at 0.1, 0.25, 0.65 and 3 m. The three default soil textural classes included in the model are those suggested by Wilson and Henderson-Sellers (1985) and are based on mineral soils (clay, loam and sand). Of particular interest to this study is the inclusion of soil water phase changes and the heat fluxes between soil layers. MOSES separates the total soil moisture (as a degree of saturation) in a given soil layer into frozen and unfrozen fractions dependant on the soil layer temperature.

A comprehensive description of the MOSES model can be found in Cox *et al.* (1999) and the Hadley Centre Technical Note 30 (Essery *et al.*, 2001).



(Cox *et al.*, 1999)

**Figure 1.2 Schematic representation of the MOSES model (showing the ‘skin’ surface temperature ( $T_*$ ), interactive surface resistance ( $r_s$ ) and soil water phase change)**

The use of MOSES to characterise hydrology and land-atmosphere interaction for a number of ecosystems, both within a GCM and as a stand alone model, has been widely documented (Blyth and Daamen, 1997; Blyth *et al.*, 1999; Blyth, 2001; 2002; Taylor *et al.*, 2002; Essery *et al.*, 2003 and Crucifix *et al.*, 2005).

However, the use of MOSES for high latitude studies is still being evaluated. These have proven to be a major test of the model since factors such as the presence of seasonally

frozen soils and spatial snow cover distribution complicate the modelling process in these regions.

Lloyd *et al.*, (1999) carried out an evaluation of MOSES for an Arctic peatland site in Kevo, Finland as part of the Land Arctic Physical Processes (LAPP) study. The evaluation of effective hydraulic parameter values for Arctic peat soils was undertaken and the ability of the model to simulate evaporation and soil temperatures was tested. In this case, realistic simulations of evaporation were obtained but the soil temperatures (especially deep soil temperatures) were not modelled well.

Harding *et al.* (2002) describe more testing of MOSES for the Kevo site in Finland. Here, measurements of the energy balance were used to test the performance of the model with respect to the exchange of energy into the top soil layer. The results showed that, although the evaporation was insensitive to the thermal properties of the top layer, the properties did influence the soil temperatures in the model.

These findings were taken further by Hall *et al.* (2003). In this study, a depth dependent parameterisation of the thermal and hydraulic properties of peat soils was incorporated into the model and evaluated. The number of soil layers in the model was also increased to examine the model performance with many, shallow soil layers. It was thought that by using many shallow soil layers with depth-dependant parameter values (such as hydraulic conductivity and soil water suction), the model performance may improved for these soils. The results of the study showed that modelled heat fluxes were close to measured values but the model did not simulate the distribution of unfrozen water distribution well. The increase in soil layers did improve the performance of the model for the soil temperatures and surface heat flux.

This chapter has provided an introduction to the potential effects of climatic change in Arctic ecosystems which form the basis of the STEPPS project. The aims and objectives of this study have been outlined and a review of current literature has provided an outline of the current thinking relevant to this thesis.

**Chapter 2**

**Methodology**

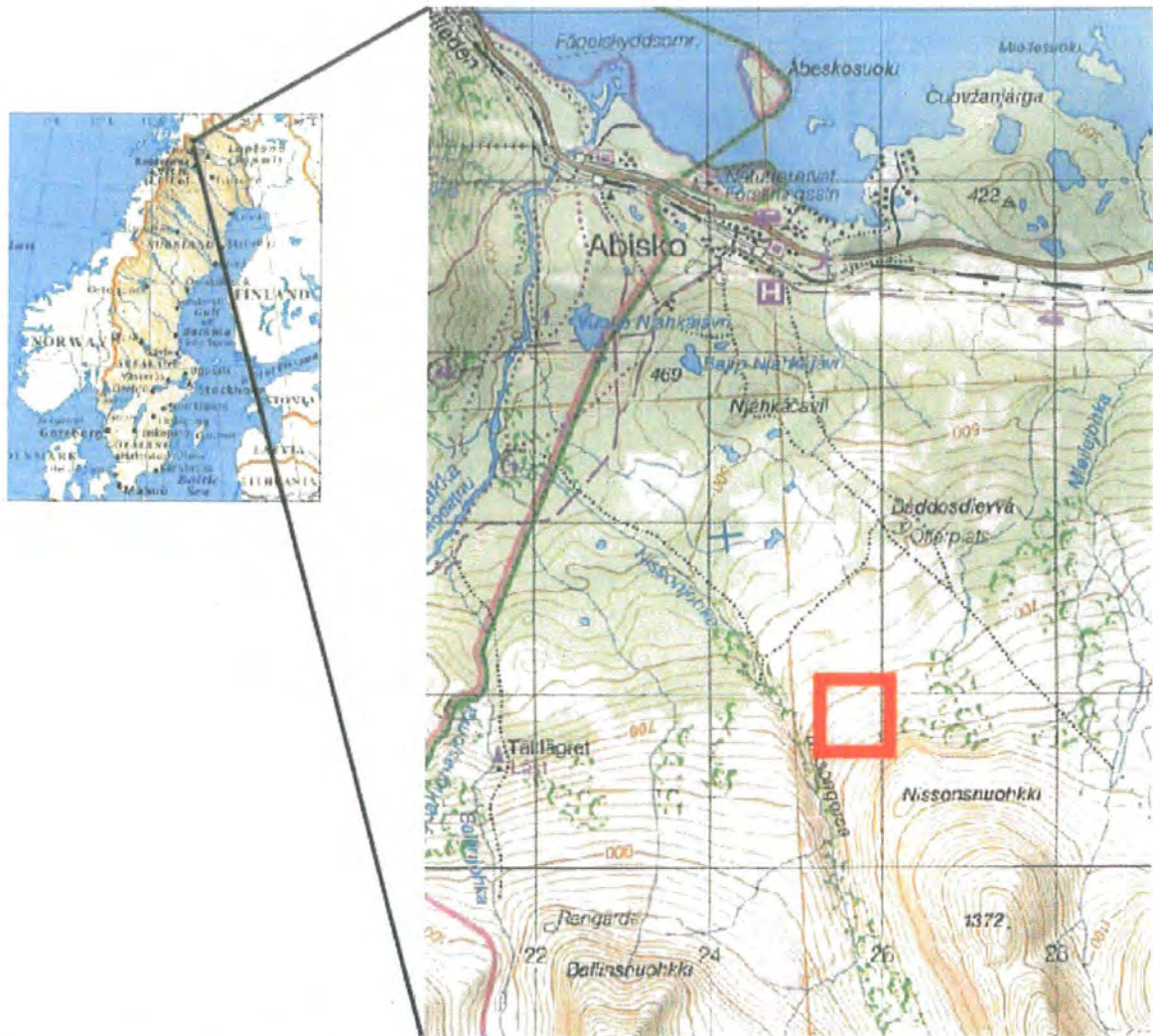
In order to investigate the project hypotheses it was necessary to undertake extensive field data collection campaigns. The first of these, in September 2002, was primarily to install automatic data collection equipment at the STEPPS fieldsite and to establish the most suitable areas for the hydrological study. Subsequent field visits were conducted in order to carry out repeated spatial soil moisture surveys, vegetation type and height surveys, a soil depth survey and a detailed topographic survey. Observations of the hydrological processes occurring in the field were also captured in photographs and notes taken throughout the season. Between September 2002 and October 2004 a total of nine field campaigns were completed.

This chapter will describe the fieldsite and outline the methods employed in the field to collect the relevant data.

## 2.1 The Study Area

The location for the field study was centred on Abisko in northern Sweden. Abisko is situated around 200km north of the Arctic Circle on the southern shore of Lake Tornetrask. Figure 2.1 shows the location of the STEPPS fieldsite and surrounding area.

The average annual temperature of the area is  $-1^{\circ}\text{C}$  with a high in June of around  $11^{\circ}\text{C}$  and low of  $-12^{\circ}\text{C}$  in January. The average annual rainfall is around 300mm. During the dark winter months (November to February), the landscape is dominated by snow and ice. During April and May the temperature increases steadily and the thawing period begins. This is followed by a short summer season (June to August) during which there is a flurry of biological activity before the days become shorter and the temperature decreases in September and October. Despite the low average annual temperature, this is not a permafrost area. The soils of the study area are seasonally frozen, the soil freeze typically occurring in early November and the thaw occurring in May.



**Figure 2.1 Map showing the location of the STEPPS study area in red (not to scale).  
(Mapquest.com) (Lantmäteriets, Fjällkarta BD6 December, 2000)**

The Abisko Scientific Research Station (ANS) was used as the base during the field campaigns. The STEPPS fieldsite is located approximately 8km south of Abisko ( $68^{\circ} 17'N$   $18^{\circ} 51'E$ ). It was accessed by ski or skidoo during the winter months and by hiking during the summer. It is between 750 m and 825 m above sea level and is a heterogeneous Tundra landscape above the treeline.

The site is comprised of a number of land cover types. *Empetrum* heath communities cover much of the site with patches of *Sedge* present in the wetter areas. *Salix* (Dwarf Willow shrub) is present in the more sheltered regions, typically occurring in the topographic hollows. The exposed ridges experience harsh conditions during the winter months being scoured by high winds and blowing snow. As a result only mosses and lichens are present on these areas.

The study area has a very distinctive topography. It is dominated by the presence of solifluction terraces which cause a series of steep sided ridges and topographic hollows to be present running parallel to the contours of the slope. These ridges and hollows are interspersed with flatter, plateau like areas which give the site a distinctly stepped appearance.

A photograph of the site can be seen in Figure 2.2. The photograph shows clearly the distinct topography of the study area, with the sheltered 'hollows' near the centre of the photograph, and exposed ridges seen in the foreground of the photograph. These features give the landscape its stepped appearance.





**Figure 2.2 Photograph showing the STEPPS fieldsite.**  
(Photograph courtesy of Anders Eriksson, ANS, September 2003)

In order to investigate the hydrology of the fieldsite and obtain data to investigate the project hypotheses, it was necessary to identify specific areas where detailed spatial and temporal measurements would be made.

#### 2.1.1 The Hydrological Study Area

The first priority of this project was to identify sub-areas within the STEPPS fieldsite that would be suitable for the hydrological investigation. In September 2002, the first visit to the study area was conducted for this purpose and a day was spent walking the site to find the landscape features deemed most appropriate. Two areas were identified that would facilitate data collection for the analysis of the hypotheses. The first was chosen for its proximity to the eddy flux correlation tower and automatic weather station (see section 2.3)

and was termed the 'Windfetch Grid'. This was an area 100 m by 200 m in size and relatively flat with an uphill slope on its southern edge. *Empetrum* heath covered most of this area, with dwarf willow scrub in a topographic hollow (running approximately east-west) to the south of the grid and exposed ridges on the northern and southern edge of the area.

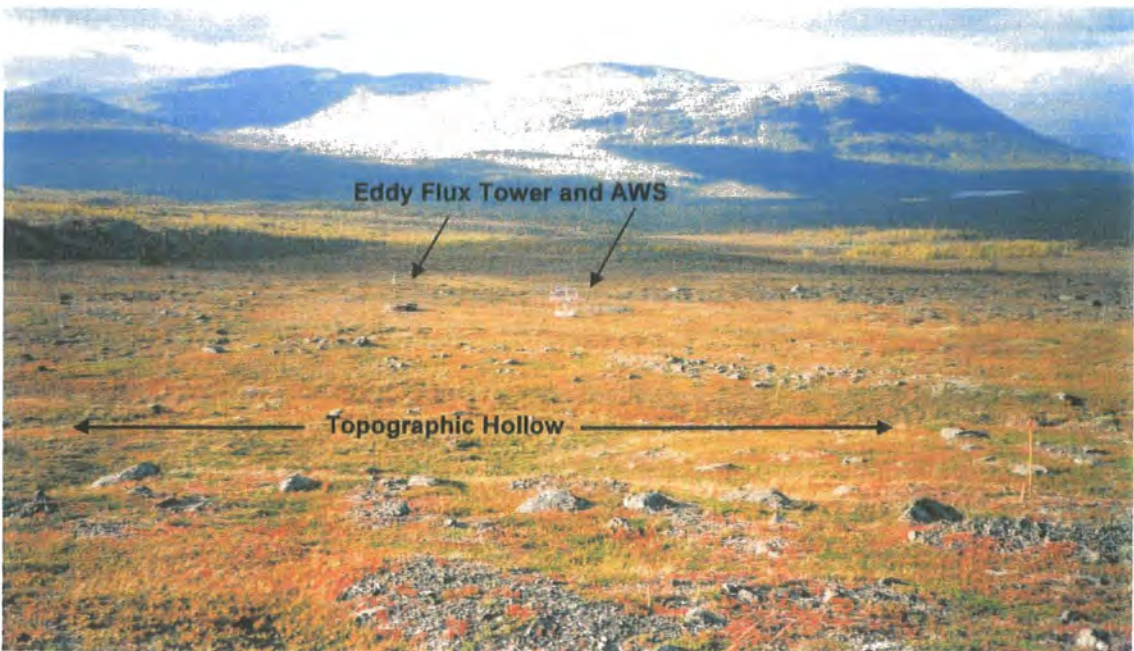
The second area chosen was a hillslope that was to the south east of the Windfetch Grid and around 50 m upslope. This site displayed a distinctly stepped appearance due to a series of steep boulder outcrops forming topographic hollows. These hollows accumulated snow in the winter months which persisted after much of the area was snow free and, once thawed, an area at the base of the outcrops appeared to be much wetter than the shallower sloping areas of the grid. The area was ideal for the investigation of the relationship between topography and soil moisture and also the effect of snowmelt on the hydrology of the hillslope. The repeated hollows and shallow sloping areas could potentially provide interesting patterns of soil moisture and, by observing these both spatially and temporally, inferences would be made as to the effect of snowmelt on the soil moisture.

The hillslope site was termed the 'Hillslope Grid' and measured 180 m long by 60 m wide. This grid was at an elevation of between 793 m and 825 m above sea level and had an approximately north facing aspect.

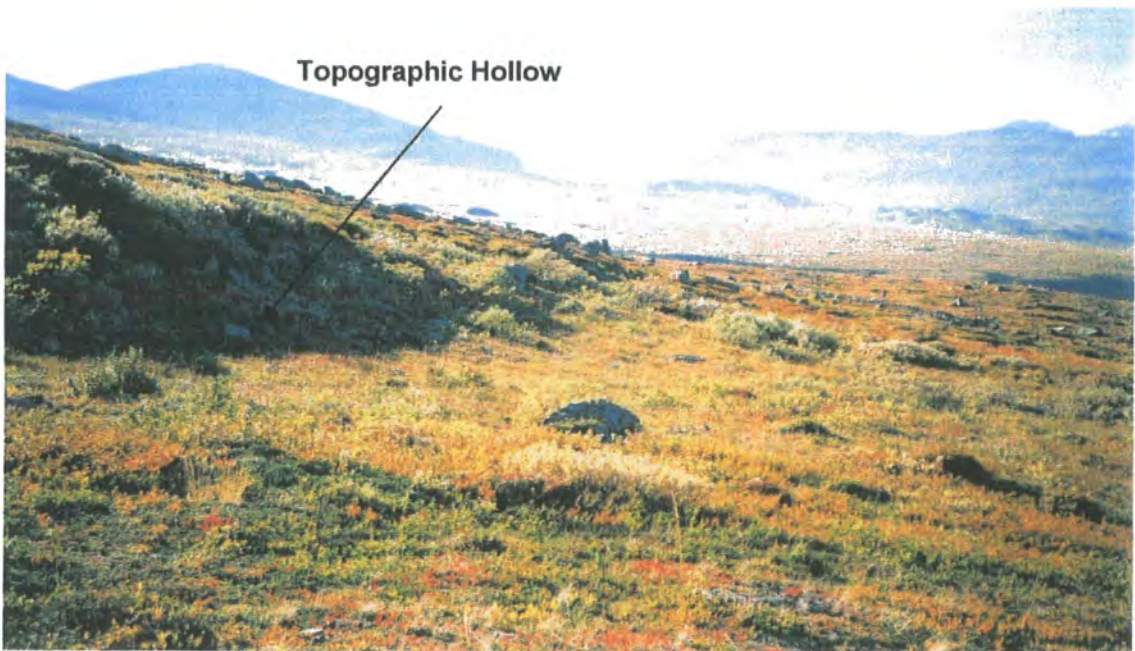
Figure 2.3 shows a view across the Windfetch Grid with a topographic hollow in the foreground. The topography dips along this southern edge of the grid before flattening out. Some exposed ridges can also be seen at the bottom of the picture. Figure 2.4 is a photograph of one of the hollows on the Hillslope Grid. Here the steep dip in topography can be seen on the left of the picture where *Salix* is dominant in this sheltered area.

Comparisons between the hillslope and non-sloping areas will be used to study the influence of topography on near surface soil moisture patterns.





**Figure 2.3** Photograph showing the features of the Windfetch Grid. View from the southern edge of the grid looking approximately north east (September 2003)



**Figure 2.4** Photograph showing the base of a topographic hollow on the Hillslope Grid. View from the eastern edge looking approximately east-west across the grid (September 2003)

## 2.2 Soil Moisture Surveys

The project relies heavily on the measurement of soil moisture across the study area in order to investigate the hypotheses. The distribution of soil moisture patterns before soil freezing and after the thaw could give an indication of soil moisture ‘memory’. Also, the affects of snowpack melt on the spatial patterns of soil moisture could only be observed by a comprehensive measurement regime that could be repeated both temporally and spatially. In order to achieve this, it was decided that grids should be used to facilitate the measurement of soil moisture at specific points across the study area for the duration of the project. By establishing these grids other physical characteristics such as soil depth, vegetation height and vegetation type could be more readily measured and spatially referenced.

### 2.2.1 Establishment of Grids

A number of factors had to be taken into consideration before the measurement sites could be established. The size of the hydrological study areas was determined primarily by examining the topography of the landscape. The scale at which measurements would be taken across the grid was influenced by the observed scale of major topographic features such as the hollows on the Hillslope Grid. A 20 m measurement interval was deemed to be too large and would exclude some possible topographic effects on soil moisture. However, a 10 m interval gave good coverage across the grid and included the topographic features at the hillslope scale. Also, since the soil moisture measurements across both grids would be compared so that the effect of topography could be assessed, it was necessary to obtain moisture measurements across both grids on the same day. This would ensure that meteorological conditions were similar at the time of moisture measurement across both study areas. Therefore it was important that the number of measurements taken could be achieved in one day. Measurement at the 10 m scale would achieve this, giving 133 grid points on the Hillslope Grid and 231 grid points on the Windfetch Grid. Also, the number of measurements over each grid would be sufficient to characterise the soil moisture distributions statistically.

A regular grid arrangement was chosen since this was the most efficient for repeated measurement in the field. Although measurement on a regular grid may not pick up linear hydrological features that may fall between grid points, the scale of measurements was appropriate for the features observed at the study area.

Having identified the locations of the hydrological study areas and their dimensions during the first field visit (September 2002), it was necessary to initially mark the positions of the grid corners with snow poles before surveying the grid. The grids were then accurately established during the May 2003 field visit using a TS315 Total Station Theodolite (Trimble Inc., U.S.). This was achieved by first establishing a baseline and then measuring a series of transects perpendicular to this baseline to form the grid.

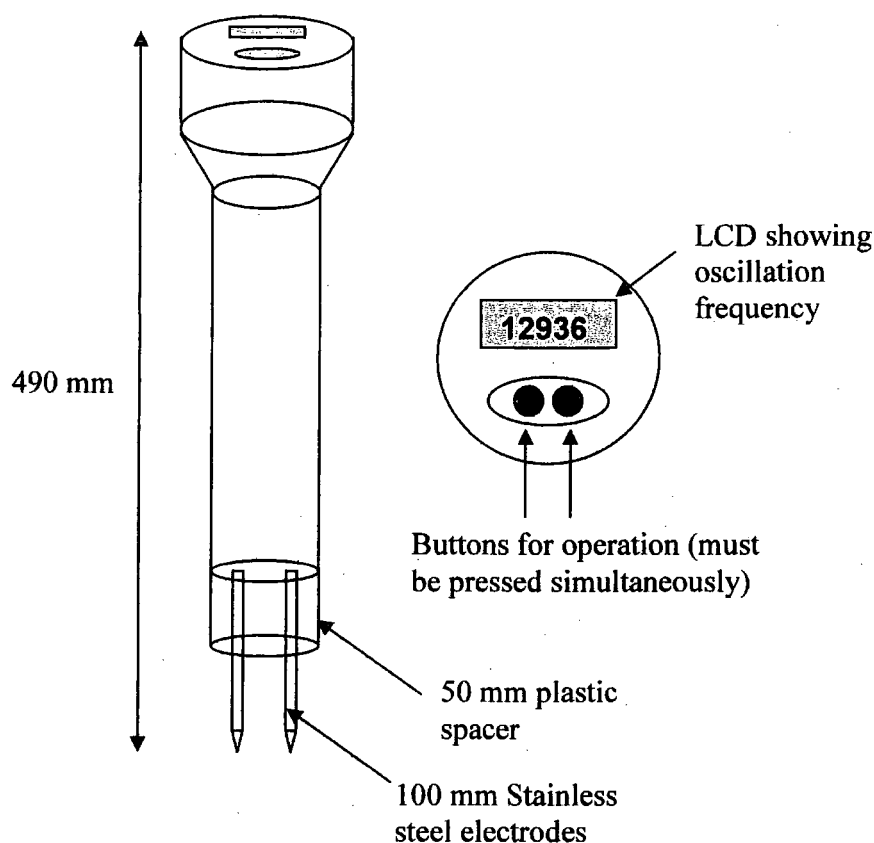
In order to mark the position of the grid points on the ground (to facilitate repeated measurements at these positions), a number of bamboo canes were cut into approximately 30 cm pegs to act as markers. Each peg was inserted into the ground at each measured grid point with minimal disturbance to the landscape.

The surveying of the grids took place between the 18<sup>th</sup> and 25<sup>th</sup> May 2003. Although the grid points could not be referenced spatially at the time, a real time kinematic survey of the points was undertaken in August and October 2004 giving accurate GPS positions of each measurement point (see section 2.4 for description of survey equipment).

Having placed pegs at all of the grid point locations it was possible to make repeated spatial soil moisture surveys using a capacitance probe.

### 2.2.2 The Surface Capacitance Insertion Probe (SCIP)

Developed at the Centre for Ecology and Hydrology (formerly the Institute of Hydrology), Wallingford, the SCIP was designed to be a portable and lightweight instrument to allow the easy acquisition of spatially distributed soil moisture measurements.



**Figure 2.5 Diagram of the Surface Capacitance Insertion Probe (SCIP) showing the instrument dimensions and characteristics**

The SCIP (Figure 2.5) measures soil moisture content indirectly by measuring the soil dielectric constant. This property can be determined by using the soil in an electrical capacitor and measuring the change in capacitance. The capacitor in the SCIP is part of an oscillator circuit that changes frequency when the capacitance changes. The rods act as the capacitor plates and, when inserted into the soil, the SCIP displays a frequency count that is related to the dielectric constant. Since the dielectric constant of the soil is related to the soil moisture content, a calibration function is used to give soil volumetric water content. Further details of the design and theory of the capacitance probe can be found in the SCIP User Guide (1994) and Robinson *et al.* (1998), the latter giving details of the calibration of the instrument.

The soil moisture measurements taken in the field were made in the top 5 cm of soil. The SCIP was operated by placing the probes vertically into undisturbed soil and pressing the two switches simultaneously. Care was taken not to stand too close to the probes and to apply the correct downward pressure on the instrument since these factors could result in measurement errors. The displayed frequency was then noted and the SCIP withdrawn from the soil.

Although the SCIP used in the field (SCIP No.6) had been calibrated for a number of mineral soils, it was necessary to calibrate the instrument for the organic soils present in the study area. This was achieved by taking soil samples from the field, drying them in an oven and determining the gravimetric water content for a number of SCIP readings. A direct relationship between soil gravimetric water content and SCIP reading was then made and the soil volumetric water content for given SCIP readings determined. A full account of the SCIP calibration methods and results can be found in Appendix A (Section A1). The variability of the SCIP was also evaluated to test the repeatability of the soil moisture measurements. This was achieved by taking four measurements at each point during a survey of the Hillslope Grid on June 2<sup>nd</sup> 2003. The variance of these points was then determined and showed that 90% of the variance values were between 0 and  $0.005 \text{ m}^3\text{m}^{-3}$ . This agrees with the experimental error found by Gardener *et al.* (1998) who suggest that a 0.005 volumetric water content error is unavoidable when using the SCIP. Full details of the evaluation of SCIP measurement error and repeatability can be found in Appendix A (Section A2).

### 2.2.3 Spatial and Temporal Soil Moisture Measurement

Soil moisture surveys were undertaken across both grids using the SCIP on a number of days during the 2003/2004 field seasons. Due to the persistence of snow and frozen ground during the thaw period, the surveys could not begin until late May of each year when the study areas were predominantly snow free (some snow persisted in the topographic hollows until early June but this affected only a few measurements on the grids). An attempt was made to measure the soil moisture across both grids on the same day, however there were

some instances when this was not possible due to weather conditions changing during the surveys. The onset of heavy rain during a survey did, on occasion, cause the survey to be abandoned since the input of water into the system could dramatically affect the soil moisture across the grids. It was therefore unwise to compare readings taken before the rain to those taken after. An effort was also made to undertake surveys on the same dates in 2004 as those in 2003 although this proved to be quite difficult due to changing weather conditions.

A total of eight surveys across the Hillslope Grid and nine across the Windfetch Grid were completed during the two years. The dates of each survey can be seen in Table 2.1.

**Table 2.1 Dates of the soil moisture surveys across the Hillslope and Windfetch Grids  
2003 to 2004**

	2003	2004
Hillslope Grid	May 28 <sup>th</sup> , June 2 <sup>nd</sup> , July 29 <sup>th</sup> , September 12 <sup>th</sup>	May 26 <sup>th</sup> , July 29 <sup>th</sup> August 10 <sup>th</sup> , September 4 <sup>th</sup>
Windfetch Grid	May 28 <sup>th</sup> , June 2 <sup>nd</sup> , July 29 <sup>th</sup> , August 10 <sup>th</sup> , September 13 <sup>th</sup>	May 26 <sup>th</sup> , July 29 <sup>th</sup> August 17 <sup>th</sup> , September 4 <sup>th</sup>

At each point across the grids the SCIP was used to take soil moisture measurements in the top 5 cm of soil. The readings were taken within 50 cm around each peg marking the grid point to avoid sensor interference from the peg. This was also necessary since measurements could not be taken in exactly the same position as the SCIP needed to be inserted into undisturbed soil for the measurement to be accurate. Therefore, every effort was made to take readings close to the peg but in undisturbed soil. Problems arose when the grid point was in a rocky or exposed ridge area since it was difficult to insert the probe into the rocky ground. Also, it is possible that the SCIP readings obtained in these rocky mineral soils may not be of the highest quality since the stones may have affected the sensor reading. In some cases the top vegetation layer needed to be carefully peeled back in order to take a soil moisture reading and then replaced. This was necessary to avoid taking a

reading through the vegetation rather than the soil and was especially important in the *Empetrum* Heath areas. Once the SCIP was inserted vertically into the soil, a reading was taken immediately and recorded in a notebook. It was important to read the instrument in this way since the observed measurement could vary dramatically if the SCIP was operated for a long period (since the instrument's circuits warm and affect the reading) or was moved away from the vertical (opening air gaps between the sensors and the soil).

### 2.3 Automatic Data Collection

At the beginning of the project (September 2002), sites were chosen for the installation of automatic datalogging equipment that would measure a variety of meteorological, soil and surface process data continually for the duration of the project.

An Automatic Weather Station (AWS) was installed on the Windfetch Grid in September 2002. This contained sensors to measure meteorological conditions such as windspeed and direction, incoming and reflected shortwave radiation, air temperature, relative humidity, atmospheric pressure, snow depth, net longwave radiation and rainfall. The sensors were connected to a CR10X datalogger and AM25T multiplexer (Campbell Scientific, Inc). Figure 2.6 shows a photograph of the Automatic Weather Station on the Windfetch Grid.

Six soil stations (SS1-6) were also installed at specific plots across the fieldsite. These soil stations were composed of a CR10X datalogger (Campbell Scientific, Inc) and AM25T multiplexer (Campbell Scientific, Inc) connected to a frost gauge (developed at the Centre for Ecology and Hydrology, Wallingford (Lloyd, 1998)) and CS616 soil water reflectometers (Campbell Scientific, Inc) at 5 cm and 15 cm depth. The frost gauge measured soil temperature at various depths through the soil profile soil moisture using thermocouples. Both the AWS and soil stations recorded sensor measurements and stored them onto SM4M storage modules (Campbell Scientific, Inc) for collection and download.





**Figure 2.6 The Automatic Weather Station (AWS1) on the Windfetch Grid**

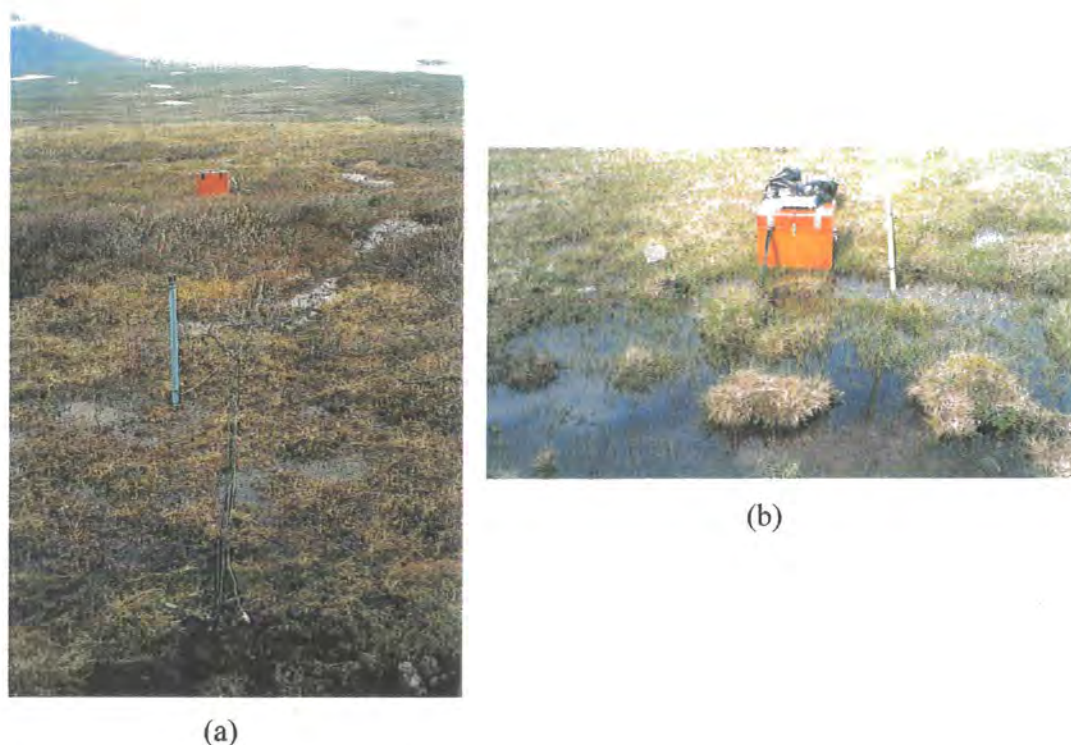
Additional sensors were installed on the Hillslope Grid and an area approximately 100 metres north east of the Windfetch Grid. The installation of CS616 Water Content Reflectometers (Campbell Scientific, Inc) on the Hillslope and Windfetch Grids allowed the continuous measurement of volumetric water content at three point locations; at *Salix* and *Empetrum* sites on the hillslope and in an area of bare ground just off the Windfetch Grid. These provided half hour averages of soil volumetric water content at 5 and 10 cm depth for the period June 2003 to September 2005 (Hillslope) and at 5, 10, 20, 50 and 70 cm depth for the period November 2003 to December 2004 (Windfetch). It should be noted that the installation of the probes at 50 cm and 70 cm depth for the Windfetch logger was problematic due to the presence of rocks and stones. Therefore, the performance of these probes is uncertain. The installation of T107 Temperature Probes (Campbell Scientific, Inc) at these sites (at the same depth as the CS616s) provided half hour averages of soil temperature. The use of a CR10X datalogger (Campbell Scientific, Inc) allowed data collection throughout the freeze and thaw periods when manual measurements were problematic. Although data from other soil stations were available, these two loggers were chosen since they provided the most complete dataset for the study period. Also, the



Windfetch logger provided data of soil moisture and temperature at soil depths not available at the other soil station sites. In addition to these data, an automatic weather station located on the Windfetch Grid (labelled AWS1 in Figure 2.8) allowed the collection of rainfall data using a CS700 Tipping Bucket Rain Gauge (Campbell Scientific, Inc.). This rain gauge was unheated and therefore only recorded rainfall and not all precipitation (rain and snow). The rainfall at this location was measured every 10 seconds and half hour totals were recorded for the duration of the study period. It should be noted that there were some periods when the automatic weather station failed to operate and some data were lost due to unavoidable technical difficulties. The rainfall data presented in Chapter 3 forms part of the most complete and reliable meteorological data record for the fieldsite during the study period and any periods of missing data have been highlighted to aid the analytical process.

Figure 2.7 shows the Hillslope Grid and Windfetch soil stations. The sensors for the hillslope logger were placed in a *Salix* plot (Figure 2.7 (a) foreground) and *Empetrum* plot (Figure 2.7 (a) background). The Windfetch logger sensors were located in a small 'stream' bed (micro-topographical feature seen in Figure 2.7 (b)) which contained standing water during the thaw period.

Field-based measurement of soil moisture using the SCIP (spatial) and CS616 Soil Water Reflectometers (point) were problematic during periods of frozen soil. The manual SCIP measurements could not be taken at points with frozen soil since the probe could not be inserted correctly into the soil. As a result, grid points with frozen soil (mainly during May and June) were omitted from the dataset. The CS616 probes do not give a true reflection of the volumetric water content in a frozen soil due to attenuation of the measurement signal; the probes measure only the unfrozen fraction of the soil water and measurement accuracy is decreased by the presence of ice. Therefore, during the periods of frozen soil, measured volumetric water contents from the CS616s should not be taken as the total amount of water in the system.



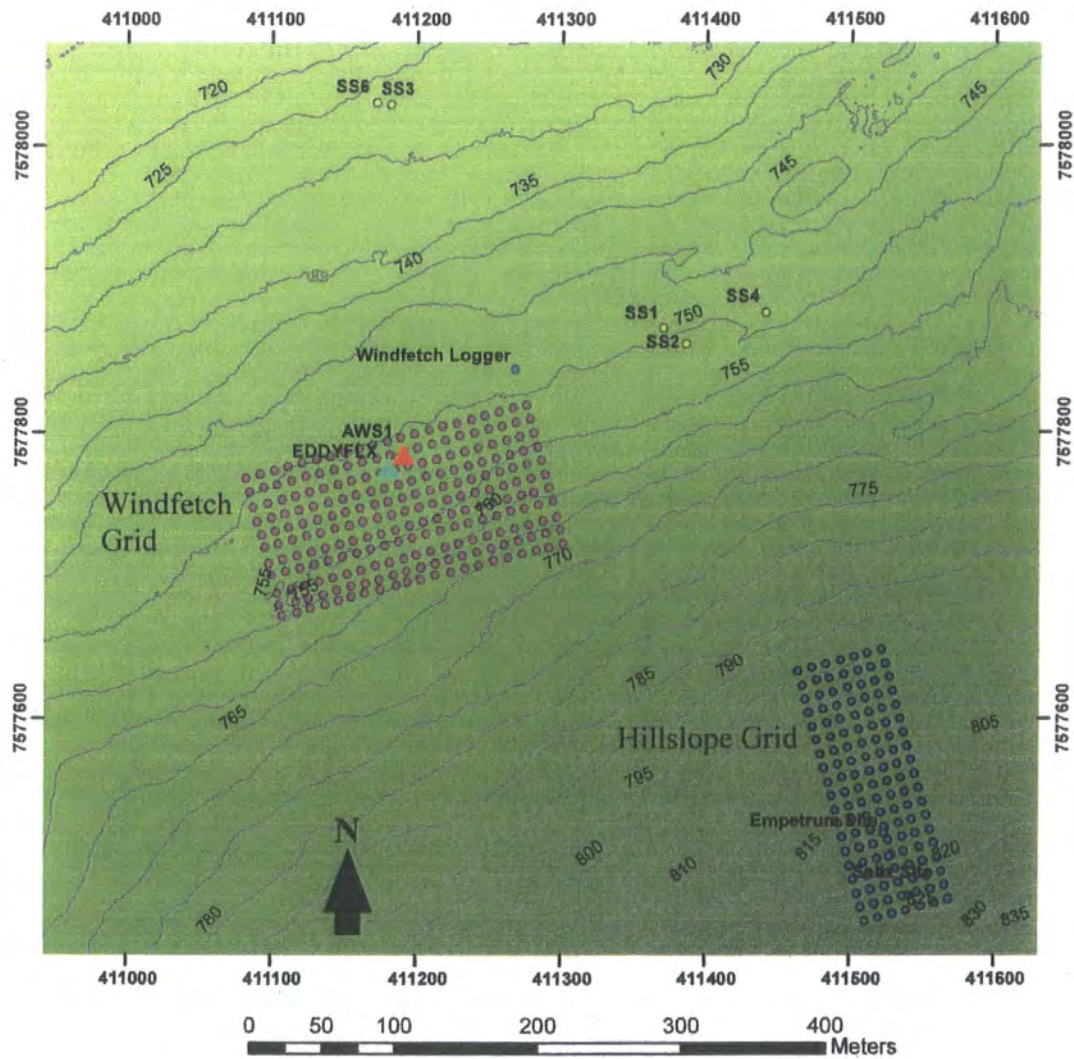
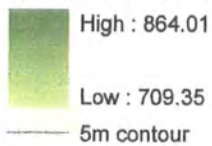
**Figure 2.7 The Hillslope soil station (a) and the Windfetch soil station (b)**

Dipwells were also installed at the logger sites (two on the Hillslope Grid and one near the second logger). An auger was used to provide access holes for the dipwell tubes to be installed. These proved to be difficult to install since the soil across the site is quite thin (70 cm maximum soil depth in places) and it was not possible to drill through the underlying rock to a reasonable depth. Therefore the dipwells were only inserted to a depth of around 30 cm on the Hillslope Grid and 85 cm at the second logger site. The dipwell water depth was measured by PDCR 1830 (Druck, UK) pressure transducers and recorded by the loggers. On examination, the data provided by these sensors were found to be not of a useable standard. This was due to the extremely problematic installation of these dipwells, mainly due to the underlying glacial deposits on the site. The presence of rocky boulders and stones under the soil horizons prevented the sinking of the dipwells to a level

approaching the groundwater level. This, coupled with the seasonal freezing of the soil water and other technical difficulties, meant that the dipwell data could not be used.

All of the loggers on the site were programmed and the sensors wired up in the standard way details of which are available in the Campbell Scientific CR10X Manual (Campbell Scientific, 2002).

Figure 2.8 shows the location of all of the automatic data collection sites and the grids used for the manual soil moisture surveys used in this study. This was produced using the ArcMap GIS software package (Environmental Systems Research Institute (ESRI), U.S).

**Elevation****LIDAR DEM****Metres (above sea level)**

Projected Coordinate System:

WGS 1984 UTM zone 34N

Datum:

WGS 1984

**Figure 2.8** Plot showing the location of measurement sites across the study area

2.4 Fieldsite Surveys

A series of field surveys were undertaken during the 2004 field season in order to capture the physical characteristics of the study area.

Dr. Robert Baxter (University of Durham) and Dr. Philip Wookey (University of Stirling) carried out soil profile investigations in July 2003. Soil pits were dug at the study area and soil profile descriptions were recorded for each dominant land cover type (*Empetrum*, *Salix*, *Sedge* and ridge areas). The descriptions included measurement of pH, Munsell classification, bulk density, loss on ignition and soil moisture content. Table 2.2 gives a summary of the soil profile data for the four land cover types. Further details of the soil profile data are given in Section 5.2.2.

**Table 2.2 Soil profile sample data (collected by Dr. Baxter and Dr. Wookey, July 2003, unpublished data, pers. comm.)**

	depth (cm)	pH	Munsell	Bulk Density	Loss on Ignition (%)
<i>Empetrum</i>	0-5	4.05	10YR 2/1	0.1045	94.84
	5-10	4.20	10YR 2/1	0.1571	87.24
	10-15	4.50	10YR 2/1	0.1477	87.91
	15-18.5	4.80	10YR 2/1	0.2754	80.28
Willow scrub ( <i>Salix</i> )	0-6	5.40	10YR 2/1	0.1564	85.31
	6-11	5.30	10YR 2/1	0.2891	67.59
	11-19	5.40	10YR 2/1	0.4644	35.38
<i>Sedge</i>	0-5	5.90	10YR 2/2	0.1436	84.38
	5-10	5.70	10YR 2/2	0.2740	76.53
	10-13	5.70	10YR 2/2	0.4103	75.10
Exposed Ridge	0-20		2.5Y 3/3		2.77

A soil depth survey was undertaken in August 2004 across both grids. A 1 m walking stick auger was inserted into the ground at each grid point and the depth to rock recorded. Four soil depth measurements were taken around each grid point and the average soil depth calculated.

A vegetation height survey was also carried out in August 2004. At each measurement point across both grids the height of vegetation was measured using a tape measure.

As there were no detailed topographic data available for the fieldsite, it was decided to undertake a topographic survey of the study areas. This was achieved using a 5700 GPS base station (Trimble, Inc) and 5800 GPS rover unit (Trimble, Inc). The base station was established at the Abisko Scientific Research Station and its location accurately fixed using baselines from the Kiruna and Tromsø GPS reference stations. Since the fieldsite was 8 km from Abisko, a TrimTalk 450S radio repeater unit (Trimble, Inc) was needed to boost the radio signal to the rover units. This was placed around 3 km from Abisko within line of sight of the study area. The rover units were used at the site to record the accurate GPS position (with cm precision) and elevation.

The real time kinematic surveying method allowed accurate GPS measurements to be made in the field in real time without needing to post process the GPS data. The rover units communicated to the controller units using Bluetooth technology. The Microsoft Windows based controller unit software allowed the user to see points as they were surveyed.

The Hillslope Grid was surveyed at 1 m resolution using the 'toposurvey' function of the rover unit. This allowed the user to walk the grid, keeping the staff as close to the ground surface as possible, while the GPS unit automatically took readings every metre. Since the user could see the points already taken on the controller unit, the entire grid could be covered at the one metre scale.

Over 11,000 readings were taken across the Hillslope Grid over a period of 7 days in August, September and October 2004. The data collected were then used to produce a detailed digital elevation model (DEM) of the hillslope using a geographical information system (ArcGIS).

The topography of the Windfetch Grid was surveyed by airborne Light Detection and Ranging (LiDAR). A remote sensing flight was carried out on the 17<sup>th</sup> July 2005 by the NERC Airborne Remote Sensing Facility (ARSF) based in Oxford, UK. This collected topographic survey points of location (easting and northing in the WGS84 coordinate system) and land surface elevation (metres above sea level) at 1.5 m resolution across the study area. This data was then used to create a DEM of the Windfetch Grid.

In order to summarise the characteristics of the field measurement campaigns, Table 2.3 shows the methods, dates of survey, sensors used, depth of measurement (where appropriate) and logging frequency for each of the surveys undertaken at the Hillslope and Windfetch sites.

The field measurement methods outlined in this section were undertaken in order to collect the data needed for the investigation of the project hypotheses. A large amount of data was collected at the STEPPS fieldsite in order to observe the physical processes operating in this remote low-Arctic environment. By using both automatic point measurements and manual spatial and temporal measurement techniques it has been possible to obtain a greater understanding of the hydrology of the area.



**Table 2.3 Summary of the Field Measurements Recorded for the Hillslope and Windfetch Sites**

	field measurement	method	date	sensor type	depth	logging frequency
<b>Hillslope Site</b>	topography	manual/point	August and October 2004	Trimble Real Time Kinematic GPS	-	1 m intervals across grid
	soil moisture	manual/spatial	see table 2.1	SCIP	top 5 cm	single measurement
		automatic/point ( <i>Empetrum</i> and <i>Salix</i> sites)	07/06/03 to 10/09/05	Campbell Scientific CS616	5 & 10 cm	every minute, 30 min averages recorded
	soil temperature	automatic/point ( <i>Empetrum</i> and <i>Salix</i> sites)	07/06/03 to 10/09/05	Campbell Scientific T107	5 & 10 cm	every minute, 30 min averages recorded
	snow depth	manual/spatial	28/02/03 and 3/05/03	marked aluminium rod	-	single measurement
	soil depth	manual/spatial	03/08/04	auger	-	single measurement
	vegetation height	manual/spatial	03/08/04	metre rule	-	single measurement
<b>Windfetch Site</b>	topography	remote sensing	17/05/05	LiDAR	-	1.5 m intervals across grid
	soil moisture	manual/spatial	see table 2.1	SCIP	top 5 cm	single measurement
		automatic/point	10/11/03 to 15/12/04	Campbell Scientific CS616	5 & 10 cm	every minute, 30 min averages recorded
	soil temperature	automatic/point	10/11/03 to 15/12/05	Campbell Scientific T107	5 & 10 cm	every minute, 30 min averages recorded
	rainfall	automatic/point	01/02/03 to 31/05/05	Campbell Scientific CS700	-	every 10 s, 30 min averages recorded
	snow depth	-	-	-	-	-
	soil depth	manual/spatial	30/08/04	auger	-	single measurement
	vegetation height	manual/spatial	30/08/04	tape measure	-	single measurement



### **Chapter 3**

**To investigate the spatial and temporal variability of soil moisture across the Tundra landscape and the processes that control this variability**

The description of the spatial and temporal soil moisture variation across the study area was complicated by the seasonal freezing and thawing cycles of the soils. The hypothesis that the spatial pattern of soil moisture across a low-Arctic Tundra heath hillslope is retained from the time the soil freezes to the subsequent thaw period was formulated in order to determine the extent to which the soil moisture changed during the winter when the soils are frozen and the effect these changes may have on the spatial distribution of soil moisture during the thaw period.

In this chapter the spatial pattern of soil moisture across the Hillslope and Windfetch Grids is described. The effects of soil freezing and thawing on these patterns are then investigated.

### 3.1 Field Methods and Data Collection

In order to determine the spatial and temporal pattern of soil moisture across the grids, a number of field methods were used. Measurements of soil volumetric water content spatially across the grids were made using the Surface Capacitance Insertion Probe (as outlined in Section 2.2.3). The dates of these surveys can be seen in Table 3.1. The SCIP measurements provided both spatial and temporal soil water content over both grids during the study period.

The Hillslope Grid logger recorded soil moisture at 5 and 10 cm depth (using CS616 soil water reflectometers (Campbell Scientific, Inc)) and soil temperature (using T107 temperature probes (Campbell Scientific, Inc)) at 5 cm depth at two locations, one in a topographic hollow and the other in an *Empetrum* area on a plateau area. The second logger recorded soil moisture and temperature at 5, 10, 20, 50 and 70 cm through the soil profile. These sensors were located in a small 'stream' bed (micro-topographical feature seen in Figure 2.7b) which contained standing water during the thaw period.

The SCIP was calibrated for the highly organic soils found at the study area. This was achieved by taking a SCIP reading in the field and then taking a soil sample at the exact

location of the reading. The moisture content of the soil sample was determined gravimetrically in the laboratory and the SCIP reading was then related to this using calibration equations. Full details of the calibration of the SCIP can be found in Appendix A (Section A.1).

Since two sensors were used for the measurement of soil water content it was necessary to evaluate the relationship between the SCIP and the CS616 soil water reflectometers (Campbell Scientific, Inc). SCIP readings taken during the spatial surveys on the 29<sup>th</sup> July 2003, 12<sup>th</sup> September 2003, 26<sup>th</sup> May 2004, 29<sup>th</sup> July 2004, 10<sup>th</sup> August 2004 and 4<sup>th</sup> September 2004 were compared with point measurements of soil moisture from the CS616s at the *Salix* and *Empetrum* sites. The results (presented in Appendix A) showed that the CS616 sensors recorded soil moisture values considerably higher than the SCIP. This difference is explained by differences in calibration of the two sensors. The SCIP was calibrated for the organic soils found at the study area whereas the CS616 sensors used a standard calibration for mineral soils. A full account of the comparison analysis can be found in Appendix A (Section A.3).

### 3.2 Analysis of Data

The data collected from the spatial SCIP surveys of both grids can be seen in Table 3.1 (below). The Hillslope Grid contained 133 measurement points and the Windfetch 231. The variation in the number of points used was due to the field survey process. In most cases this was due to the presence of snow or frozen ground which prevented measurements being taken at these grid points. The measurements taken over the Windfetch Grid on the 28<sup>th</sup> May 2003 are missing two grid transects which could not be measured in the time available on that day.

The mean volumetric water content for the Hillslope Grid was greatest in May (for both 2003 and 2004). The lowest mean volumetric water content occurred in July 2003 ( $0.19 \text{ m}^3 \text{ m}^{-3}$ ). Standard deviations were highest during the wettest months (May and June 2003 and May 2004) dropping significantly later in the season. All measurements showed a large

range in volumetric water content, the highest being  $0.55 \text{ m}^3 \text{ m}^{-3}$  in May 2004 on the Hillslope Grid and the lowest being  $0.27 \text{ m}^3 \text{ m}^{-3}$  in August 2003 on the Windfetch Grid.

**Table 3.1 Mean, standard deviation, range and number of data points from the SCIP survey data for the Hillslope and Windfetch Grids**

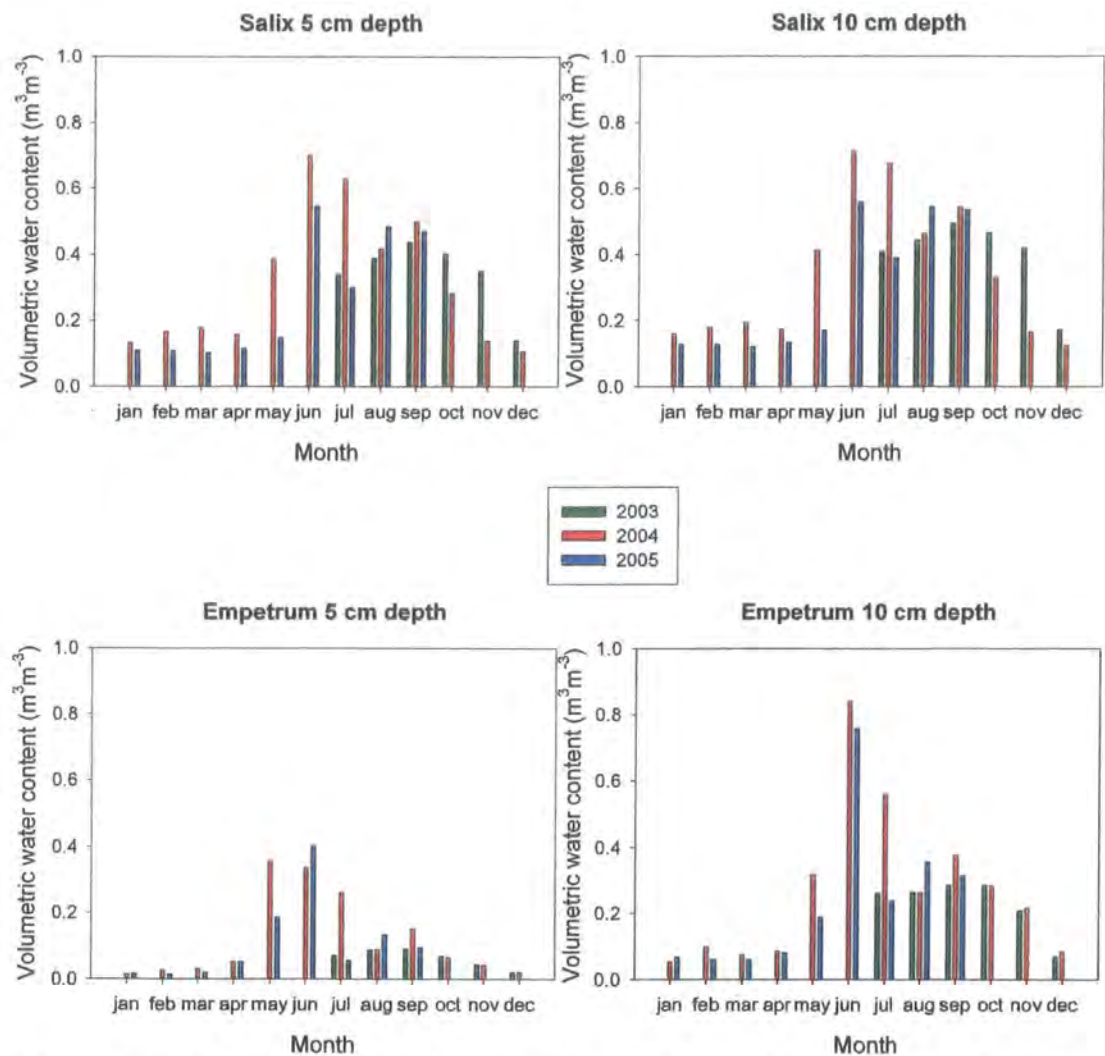
	Hillslope Grid								
	2003					2004			
	28-May	02-Jun	29-Jul	10-Aug	12-Sep	26-May	29-Jul	10-Aug	04-Sep
mean (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.33	0.30	0.19		0.21	0.34	0.27	0.22	0.22
std dev (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.13	0.10	0.06	No data	0.07	0.12	0.10	0.08	0.07
range (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.48	0.42	0.28		0.33	0.55	0.41	0.37	0.32
number of points	130	131	133		133	122	133	133	133

	Windfetch Grid								
	2003					2004			
	28-May	02-Jun	29-Jul	10-Aug	13-Sep	26-May	29-Jul	17-Aug	04-Sep
mean (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.35	0.28	0.18	0.17	0.19	0.30	0.28	0.25	0.21
std dev (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.12	0.12	0.05	0.05	0.06	0.11	0.11	0.06	0.07
range (vwc $\text{m}^3 \text{ m}^{-3}$ )	0.49	0.44	0.30	0.27	0.35	0.46	0.47	0.37	0.39
number of points	182	226	231	229	231	219	229	231	231

Figure 3.1 shows the monthly average unfrozen volumetric water content for the *Salix* and *Empetrum* sites on the hillslope derived from CS616 water content reflectometer data. The logged data were averaged over each calendar month for the duration of the study (June 2003 to September 2005) and graphed to show the monthly and annual variation in soil volumetric water content for the two sites at both 5 cm and 10 cm soil depth.

At these sites, the soil thaws between April and June and freezes between October and November (the timing of the freeze and thaw cycles will be described later in this chapter).



**Figure 3.1 Monthly Average Volumetric Water Content ( $\text{m}^3 \text{m}^{-3}$ ) for *Salix* and *Empetrum* sites on the Hillslope Grid during the Study Period (June 2003 - September 2005)**

At the *Salix* site it is apparent that there is very little difference between the monthly averages at the two depths.

During 2003, the monthly average soil volumetric water content at both depths rises steadily between July and September, before falling until the freeze in mid November. In 2004, a large increase between April and May (approx.  $0.23 \text{ m}^3 \text{m}^{-3}$ ) was observed (due to

the thaw) followed by a larger increase between May and June (approx.  $0.30 \text{ m}^3 \text{ m}^{-3}$ ). The averages then decrease during July and August, rise slightly in September then fall steadily until the freeze, as seen in 2003. During 2005, the large increase in average volumetric water content in the spring occurs later than 2004 (between May and June). A decrease is observed in July (slightly larger at 5 cm depth than 10 cm depth) before rising in August and September.

The largest average volumetric water contents for the *Salix* site were observed during June and July 2004 ( $0.70$  and  $0.62 \text{ m}^3 \text{ m}^{-3}$  at 5 cm depth and  $0.71$  and  $0.68 \text{ m}^3 \text{ m}^{-3}$  at 10 cm depth).

The *Empetrum* site shows a similar pattern of average soil volumetric water content to the *Salix* site during the study period. During 2003, an increase between July and September is observed, followed by a decrease during November and December. The actual values are considerably lower than for the *Salix* site. In 2004, an increase is observed in May but, at 5 cm depth, this is followed by a decrease in June. This is not the case at 10 cm depth, where a large increase is observed between May and June ( $0.52 \text{ m}^3 \text{ m}^{-3}$ ). The monthly pattern is then similar to the *Salix* site, but with lower average volumetric water content. The 2005 data show a large increase during April, May and June. The largest increase for all years and both sites is observed at the *Empetrum* site at 10 cm depth between May and June ( $0.57 \text{ m}^3 \text{ m}^{-3}$ ). The pattern is then similar to the *Salix* site between June and September.

At the *Empetrum* site at 5 cm depth, the water content does not rise above  $0.4 \text{ m}^3 \text{ m}^{-3}$  during the study period. However, at 10 cm depth, the highest average soil water contents for both sites were observed in June 2004 and 2005 ( $0.84$  and  $0.76 \text{ m}^3 \text{ m}^{-3}$  respectively).

A difference in average volumetric water content has been observed between the *Salix* and *Empetrum* sites which may be due to a number of factors including soil depth, vegetation type, vegetation height and topography. The influence of these factors on soil moisture at the study site will be explored further in later chapters.

Further exploration of the pattern of soil water content through time were carried out using data collected during the spatial surveys of the Hillslope and Windfetch Grids. Initially, these data were used to produce frequency histograms for both grids between May 2003 and September 2004 (Figures 3.2 and 3.3) using bins of  $0.02 \text{ m}^3 \text{ m}^{-3}$ .

In May the distributions for both grids and both years are all approximately symmetric with only a slight negative skew. In 2003, the Hillslope Grid distributions are all slightly positively skewed and in July and September the highest frequencies occur towards the dry end of the distributions. This is also seen on the Windfetch Grid during July, August and September 2003 (the September distribution being positively skewed). During 2004, both grids show a positive skew and higher frequencies towards the dry end of the distributions which become more pronounced with time from July to September.

The histograms indicate that as the season progresses from the thaw period in May and June to the end of the summer in September, the wetter conditions give way to much lower soil moisture conditions.

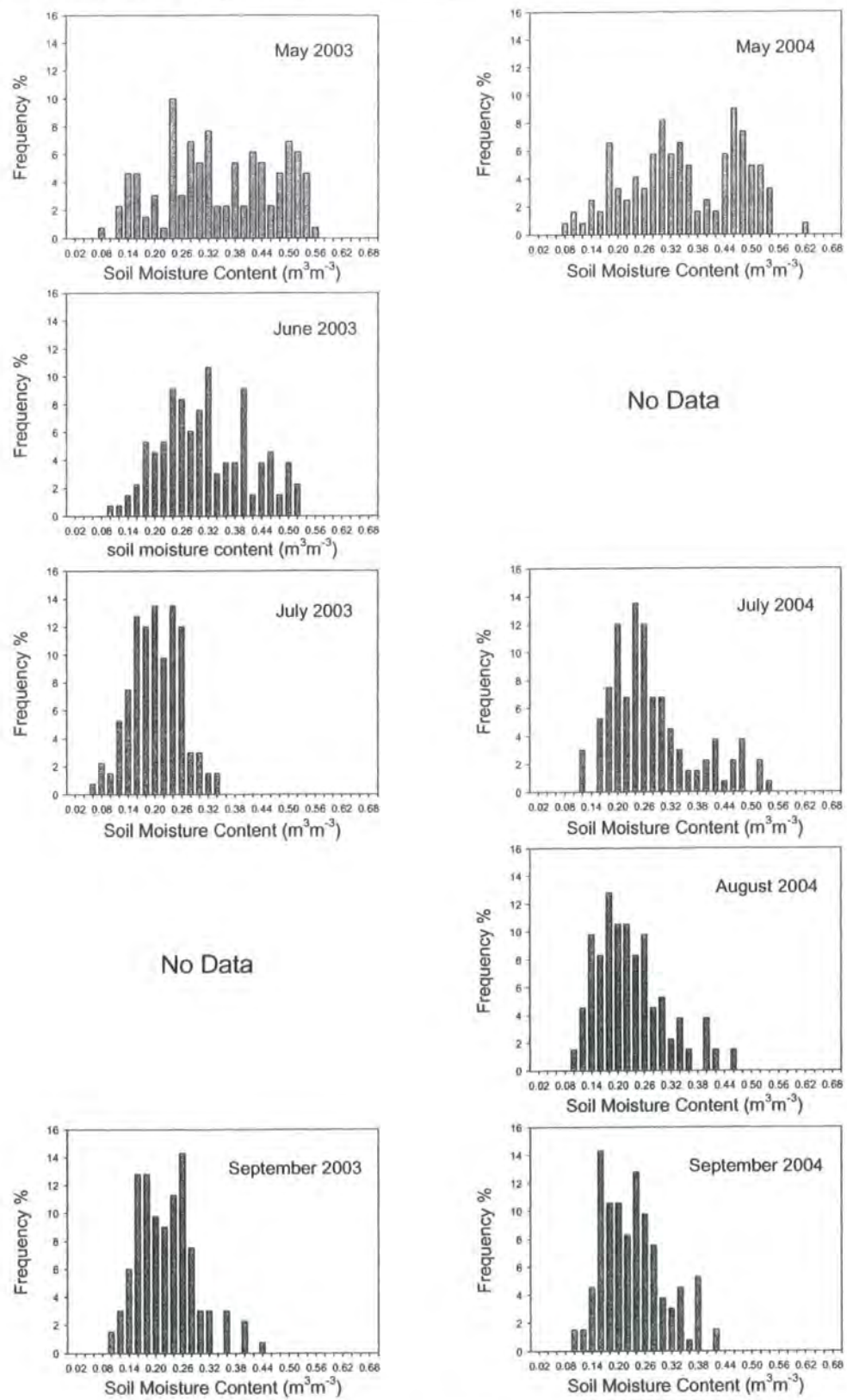


Figure 3.2 Frequency histograms of the Hillslope Grid for each spatial soil moisture survey



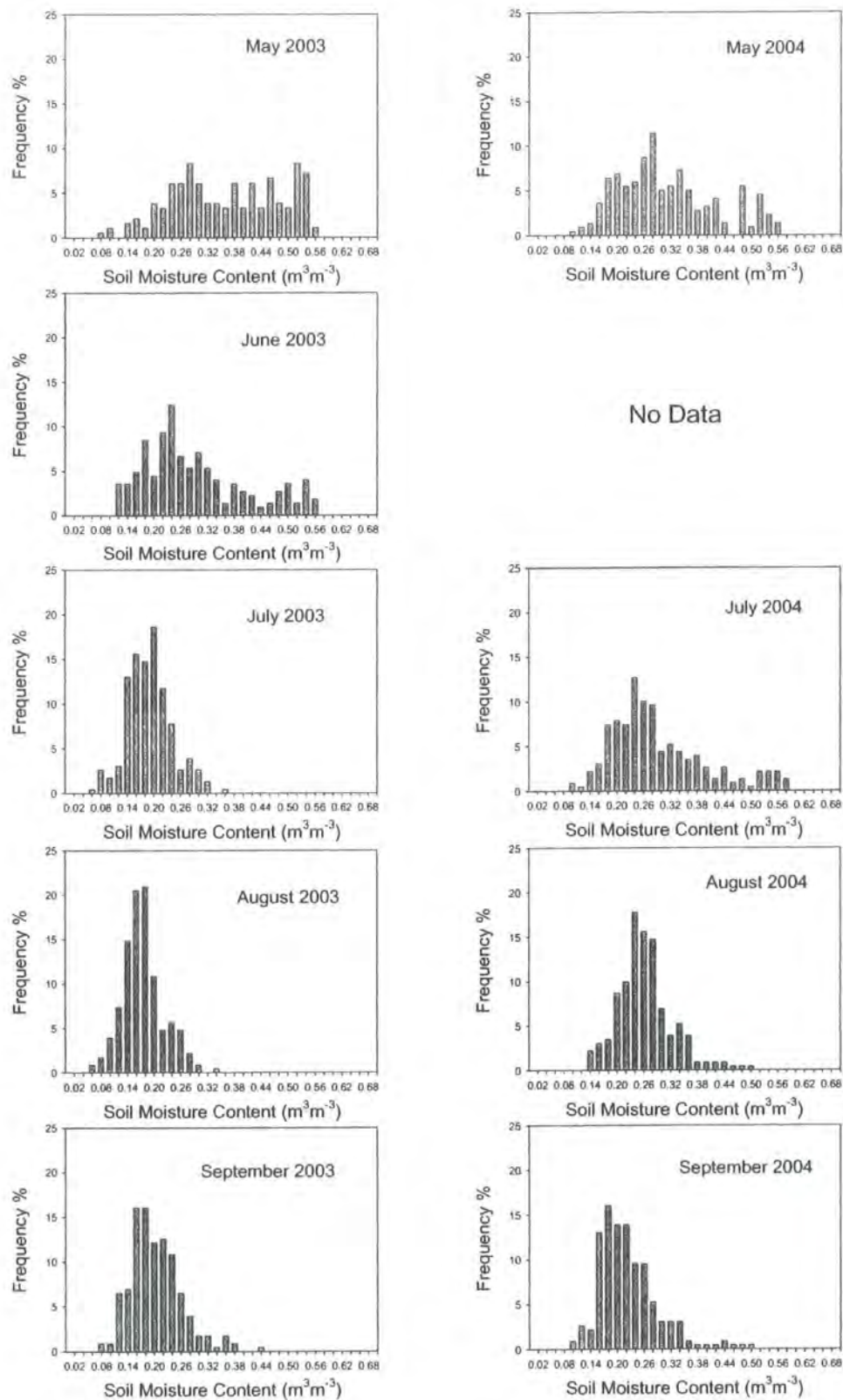


Figure 3.3 Frequency histograms of the Windfetch Grid for each spatial soil moisture survey

### 3.2.1 Variogram Production

A useful method for describing the spatial pattern of soil moisture was the production of variograms for the SCIP survey datasets. This statistical method gives an indication of the degree to which points separated by short distances show similar values compared to those further away (spatial autocorrelation). This method is also used for interpolation methods (such as kriging) in order to display spatial patterns of soil moisture derived from point measurements. The production of directional variograms is useful in order to investigate any preferred directionality (anisotropy) of soil moisture patterns within the datasets.

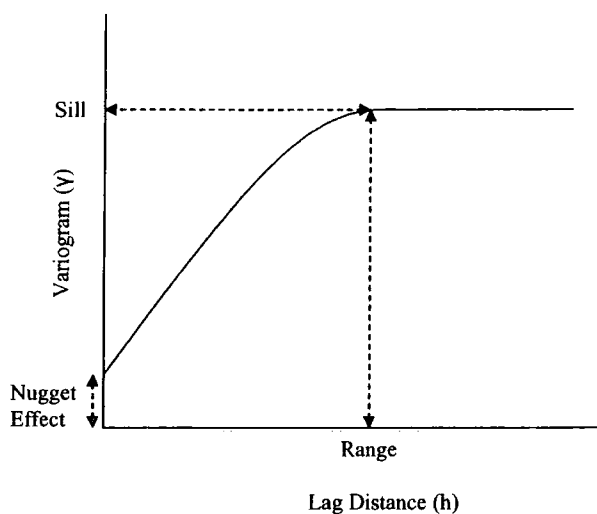
The variogram is defined as being half the average square difference between the paired data values:

$$\gamma(h) = \frac{1}{2N(h)} \sum_{(i,j)} (\theta_i - \theta_j)^2$$

(Equation 3.1)

(Western *et al.*, 1998a)

There are three main characteristics of the variogram that should be considered (shown in Figure 3.4). The range is the distance at which the variogram flattens out and reaches a plateau. This represents the maximum distance over which spatial correlation is present. The sill is the plateau and represents the variance between two points separated by a large distance. The nugget effect is the variance between two points that are spatially close together and is caused by measurement error or spatial variability at a scale smaller than the measurement scale (Isaaks and Srivastava, 1989, Loague, 1992, Western *et al.*, 1998a, Anctil *et al.*, 2002, Western *et al.*, 2003, Petrone *et al.*, 2004).

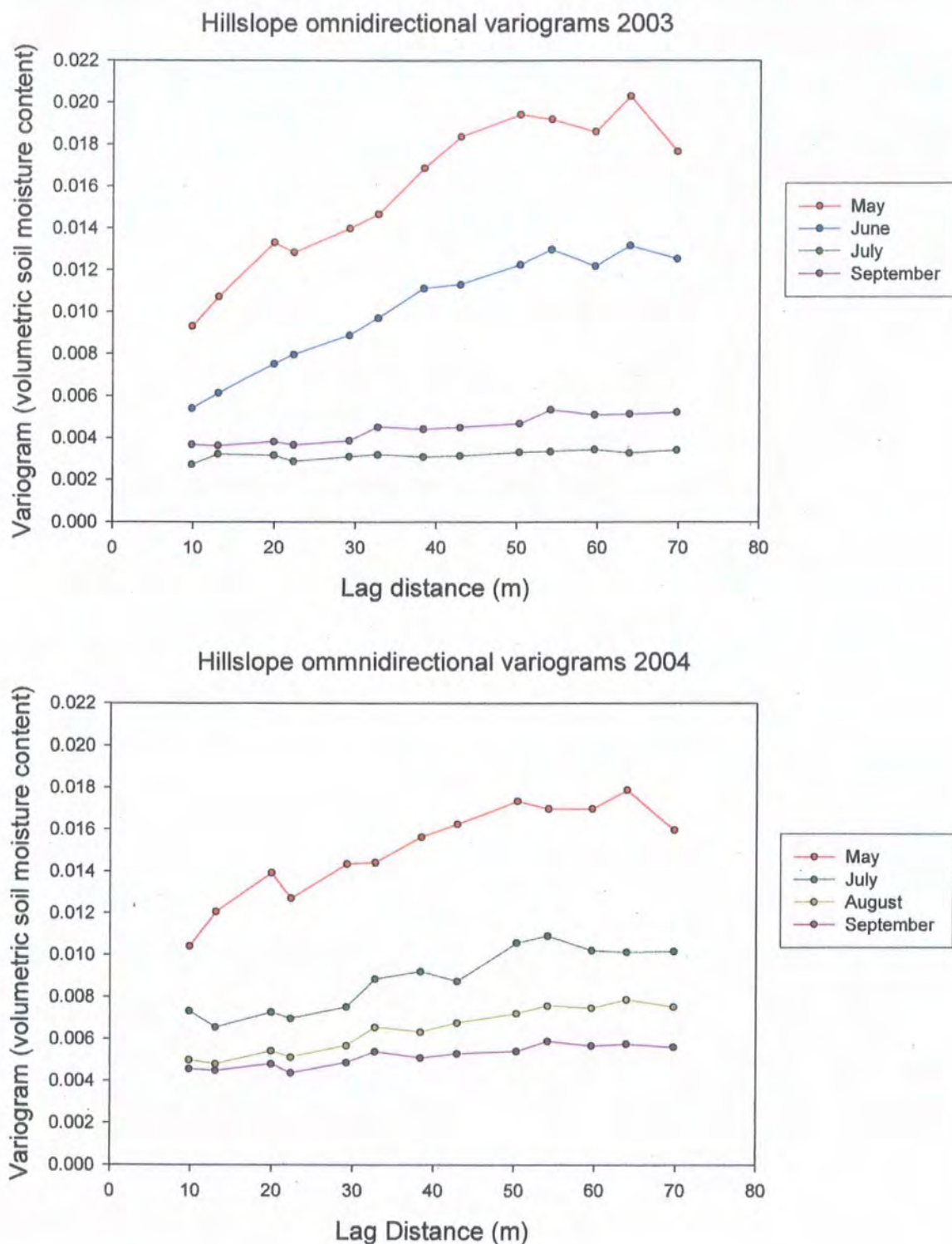


**Figure 3.4 Characteristics of a typical variogram (after Isaaks and Srivastava, 1989)**

Omnidirectional variograms were produced using the Surfer 7 software package (Golden Software, Inc., 1999) for the Hillslope and Windfetch Grids using data from the spatial soil moisture surveys. A lag bin of 5.14 m was used for the Hillslope Grid data, with a maximum lag distance of 72 m and using 14 lags. The Windfetch Grid data was analysed using a lag bin of 5.11 m, a maximum lag of 87 m and using 17 lags.

Figure 3.5 shows omnidirectional variograms for the Hillslope Grid for 2003 and 2004. The variograms show a distinct seasonal trend. The data for May 2003 show a well defined sill at around 0.019 (close to the variance of 0.016) with a range of approximately 50 m. The June variogram has a lower sill (approximately 0.012) with a range of 55 m. There is then a distinct drop in the sill values for July and September (approx. 0.003 and 0.005 respectively). This may be due to the lower soil moisture values for these months and the fact that the moisture is distributed more evenly across the grid. In May 2004, a similar pattern is displayed as that in 2003. The variogram for July 2004 shows a higher sill (approx. 0.010) compared to the 2003 data. This is a result of the higher soil moisture for this survey.

The high sills during the wetter periods may be due to the redistribution of soil moisture caused by topography. This increases variability during the wet periods.



**Figure 3.5 Omnidirectional variograms for the Hillslope Grid derived from spatial soil moisture survey data**

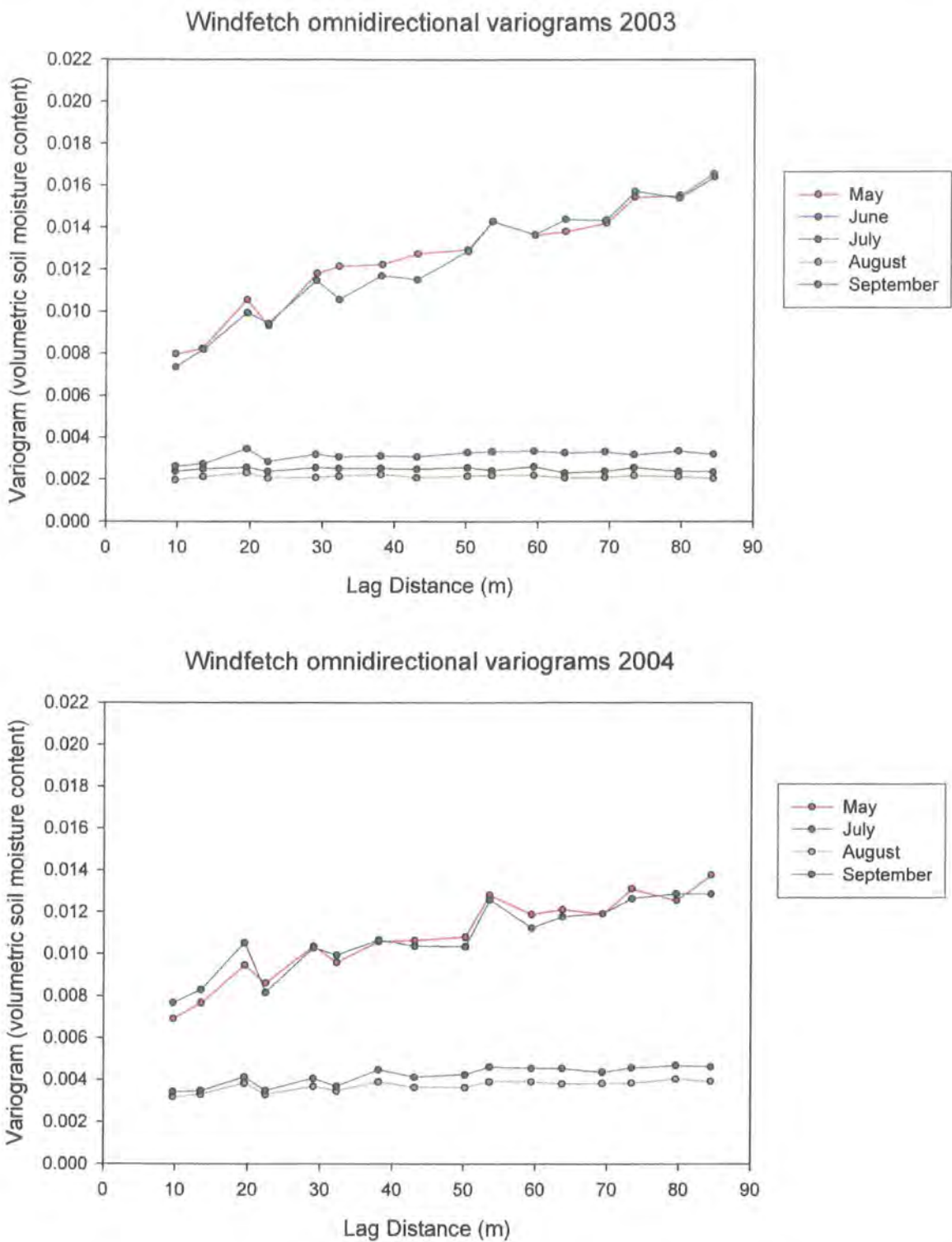


Figure 3.6 shows omnidirectional variograms for the Windfetch Grid for 2003 and 2004. Again a seasonal trend can be inferred from the variograms but it is difficult to distinguish a distinct sill for May and June 2003. The July, August and September variograms all have sills close to their variances (0.003, 0.002 and 0.003 respectively). The 2004 variograms for May and July are very similar and show a sill of approximately 0.012 and a range of approximately 55 m. Again, the August and September variograms show a sill similar to their variance.

### 3.2.2 Anisotropy

In order to investigate any preferred directional variability in soil moisture, directional variograms were produced for the Hillslope and Windfetch Grids (from the spatial soil moisture surveys of May and September 2004). In all cases, an angular tolerance of  $\pm 30^\circ$  was selected and the north-south and east-west directions were plotted using the Surfer 7 software. The corresponding omnidirectional variograms were also plotted to show any variation in the directional plots.

Figure 3.7 shows directional variograms for the Hillslope Grid. The hillslope variograms for May 2004 show higher variability in the east-west direction (a higher sill) than in the north-south direction. The directional plots are significantly different to the omnidirectional variogram suggesting that the structure is anisotropic. This may be due to the topography of the hillslope. Since the hillslope is north-facing, drainage patterns tend to occur in the north-south direction and there is maximum continuity in this direction during wet periods.

Figure 3.8 shows directional variograms for the Windfetch Grid. The directional plots for May 2004 are much more similar to the omnidirectional variogram. Again, the flatter topography of the Windfetch Grid may explain this. The directional and omnidirectional variograms for both grids for September 2004 show very little variation and can be considered to be isotropic (no directional variability). The effect of topography on soil moisture distribution will be explored further in Chapter 4.



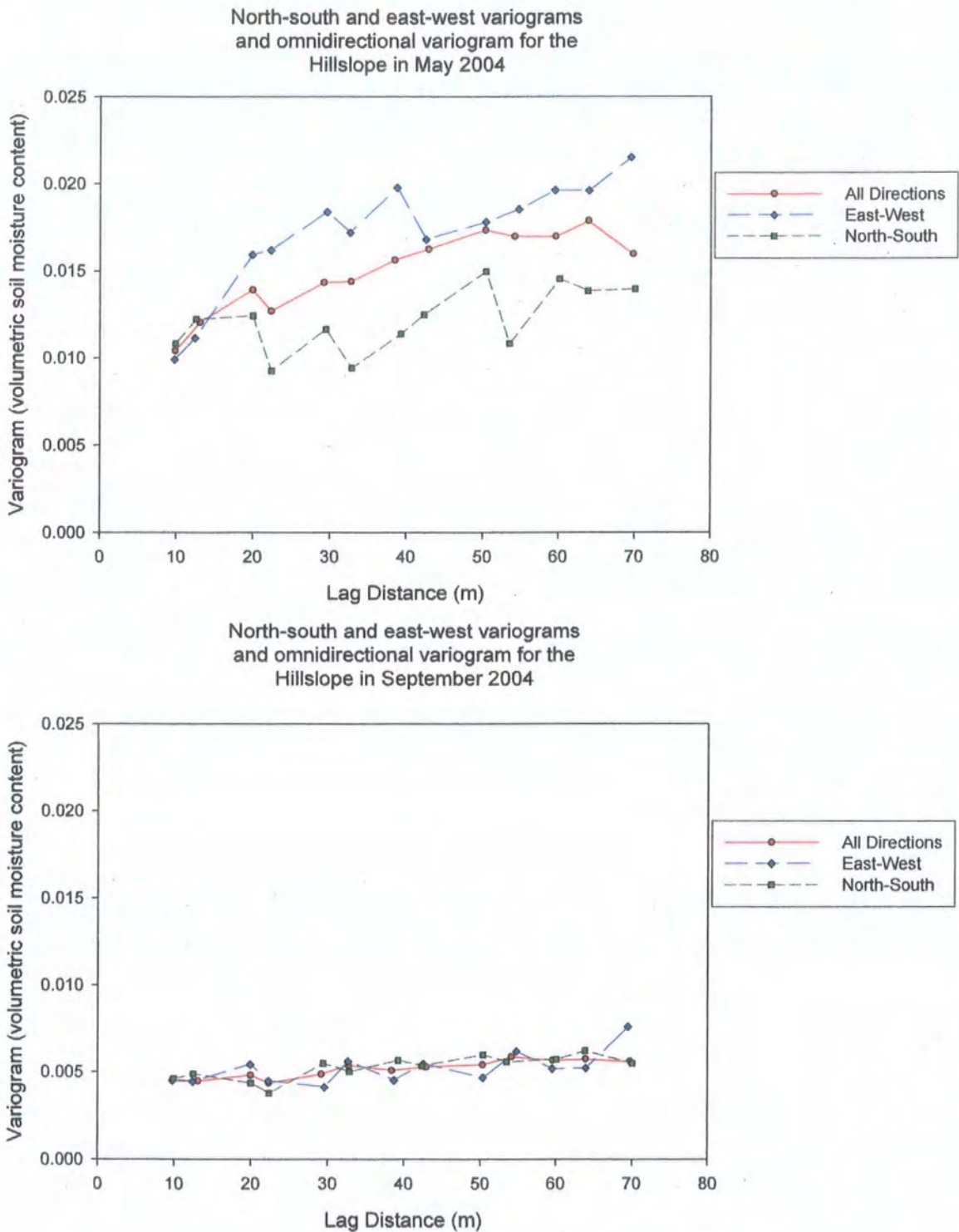


Figure 3.7 Directional and omnidirectional variograms of the Hillslope Grid for May and September 2004

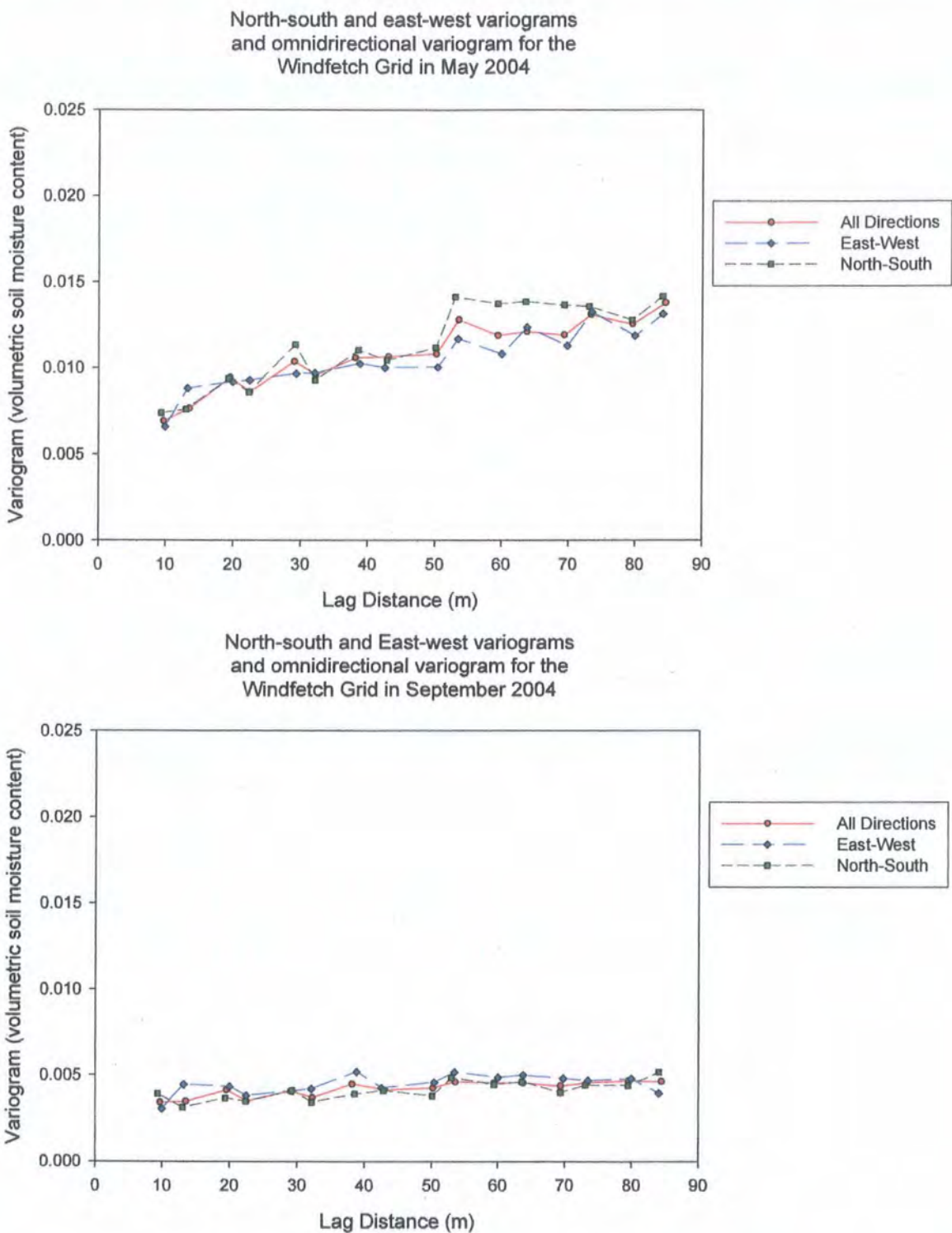


Figure 3.8 Directional and omnidirectional variograms of the Windfetch Grid for May and September 2004



### 3.2.3 Variogram Modelling

Having produced the omnidirectional variograms for both grids it was then necessary to apply a variogram model that best fitted the data. This was undertaken since the models could then be used for the interpolation process to give a representation of the soil moisture patterns across the grids.

The Surfer 7 software (Golden Software, Inc., 1999) provides a choice of models that can be used to give a best fit for the data, including spherical and exponential models.

The spherical model uses the following equation:

$$\gamma(h) = \begin{cases} C[1.5h - 0.5h^3] & h < 1 \\ C & h \geq 1 \end{cases}$$

(Equation 3.2)

(Golden Software, Inc., 1999)

where  $\gamma(h)$  is the semi-variogram,  $C$  is the scale for the structured component of the variogram and  $h$  is the anisotropically rescaled, relative separation distance (Golden Software, Inc., 1999). This has a form similar to that shown in Figure 3.4 with an initial linear relationship at short distances and leveling off to a well defined sill.

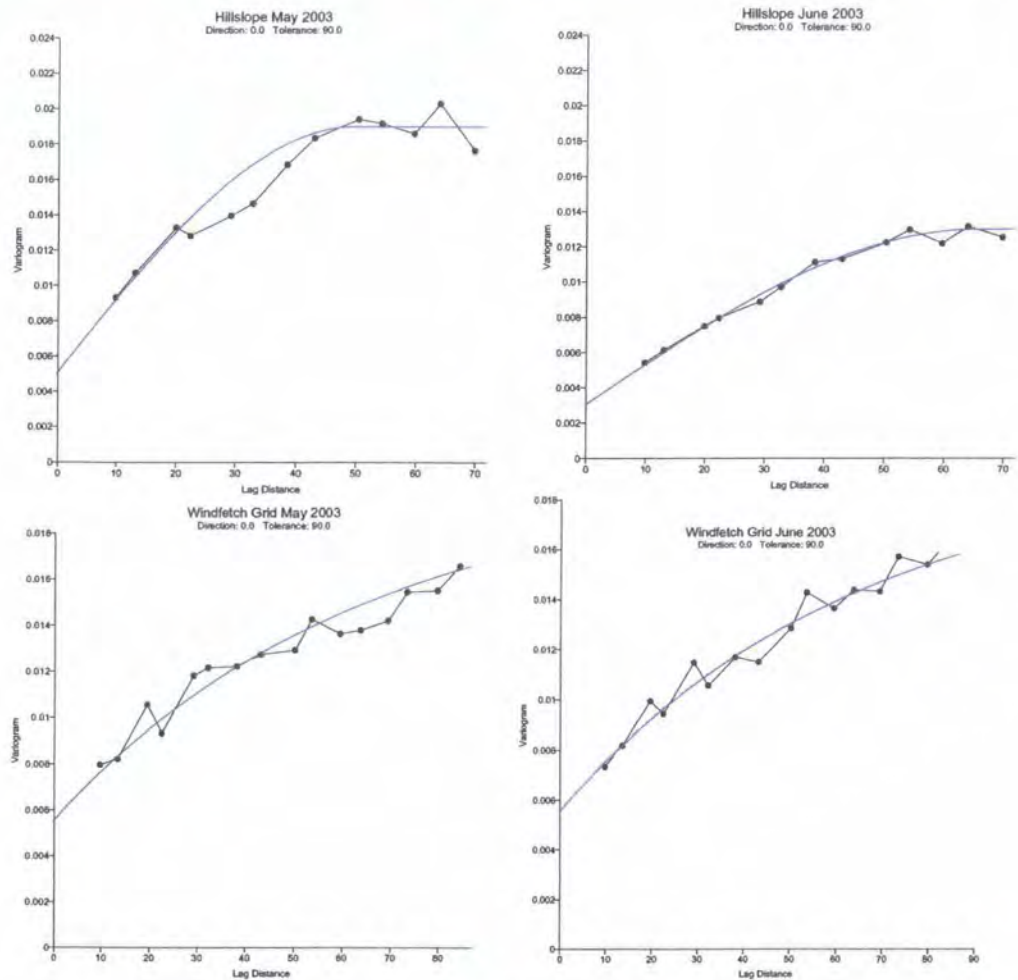
The exponential model uses the following equation:

$$\gamma(h) = C[1 - \exp(-h/a)]$$

(Golden Software, Inc., 1999)

where  $a$  is the theoretical range. This model has a linear relationship at short distances but rises steeply and flattens out to the sill more gradually (Isaaks and Srivastava, 1989).

On visual inspection of the omnidirectional variograms produced for the Hillslope and Windfetch Grids, the models that provided a best fit were the spherical model (with nugget) for the hillslope data and an exponential model (with nugget) for the Windfetch data.



**Figure 3.9 Examples of variogram modelling for the Hillslope Grid (spherical model) and Windfetch Grid (exponential model) for May and June 2003**

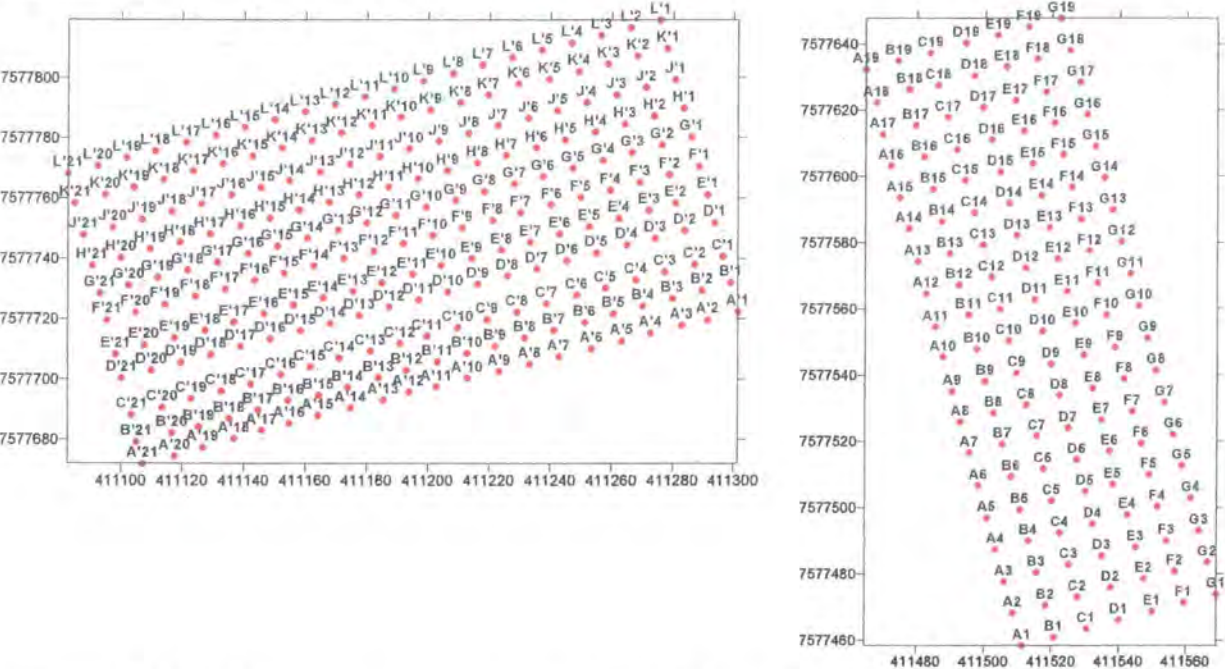
Table 3.2 shows the variogram nugget values for the modelled omnidirectional variograms for the Hillslope and Windfetch sites for each of the spatial soil moisture surveys. The modelled variograms for the hillslope and Windfetch soil moisture data show nugget values ranging from 0.006 to 0.0015. The highest nugget values occur during the spring months and decrease through the summer months. The main cause of the nugget effect is

measurement error. The nugget values observed in this study are consistent with the measurement error associated with the use of the SCIP (Appendix A, Section A.2).

**Table 3.2 Variogram nugget values for the modeled omnidirectional variograms for the Hillslope (spherical model) and Windfetch (exponential model) sites for the study period**

	<b>Hillslope Grid</b>	<b>Windfetch Grid</b>
<b>May 2003</b>	0.005	0.0055
<b>Jun 2003</b>	0.003	0.0055
<b>Jul 2003</b>	0.0015	0.002
<b>Aug 2003</b>	-	0.0015
<b>Sep 2003</b>	0.003	0.0015
<b>May 2004</b>	0.006	0.0055
<b>Jun 2004</b>	-	-
<b>Jul 2004</b>	0.005	0.005
<b>Aug 2004</b>	0.003	0.002
<b>Sep 2004</b>	0.0035	0.0015

Once the models were estimated for all the spatial survey data, an interpolated surface could be produced for the grids using the Surfer 7 software. A kriging method was applied that used the variogram model for each survey for interpolation. Kriging is an interpolation method that uses weighted linear combinations of the measured data (Isaaks and Srivastava, 1989) to estimate values between data points. The interpolation process also removes the measurement error associated with the field measurements. The results of the kriging process are maps of soil moisture distribution across the grids.

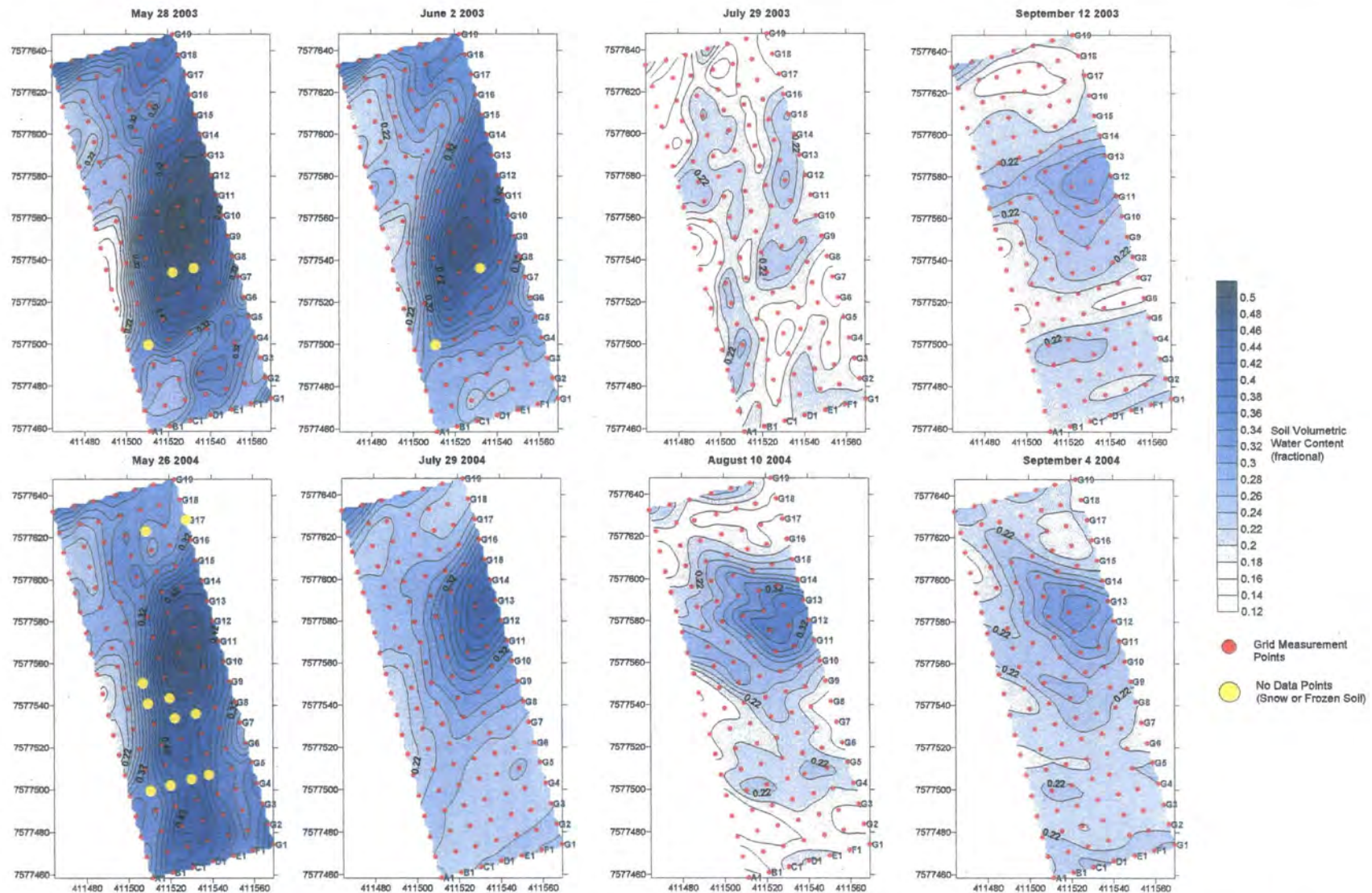


**Figure 3.10** Grid point naming conventions for the Windfetch Grid (above) and the Hillslope Grid (right)

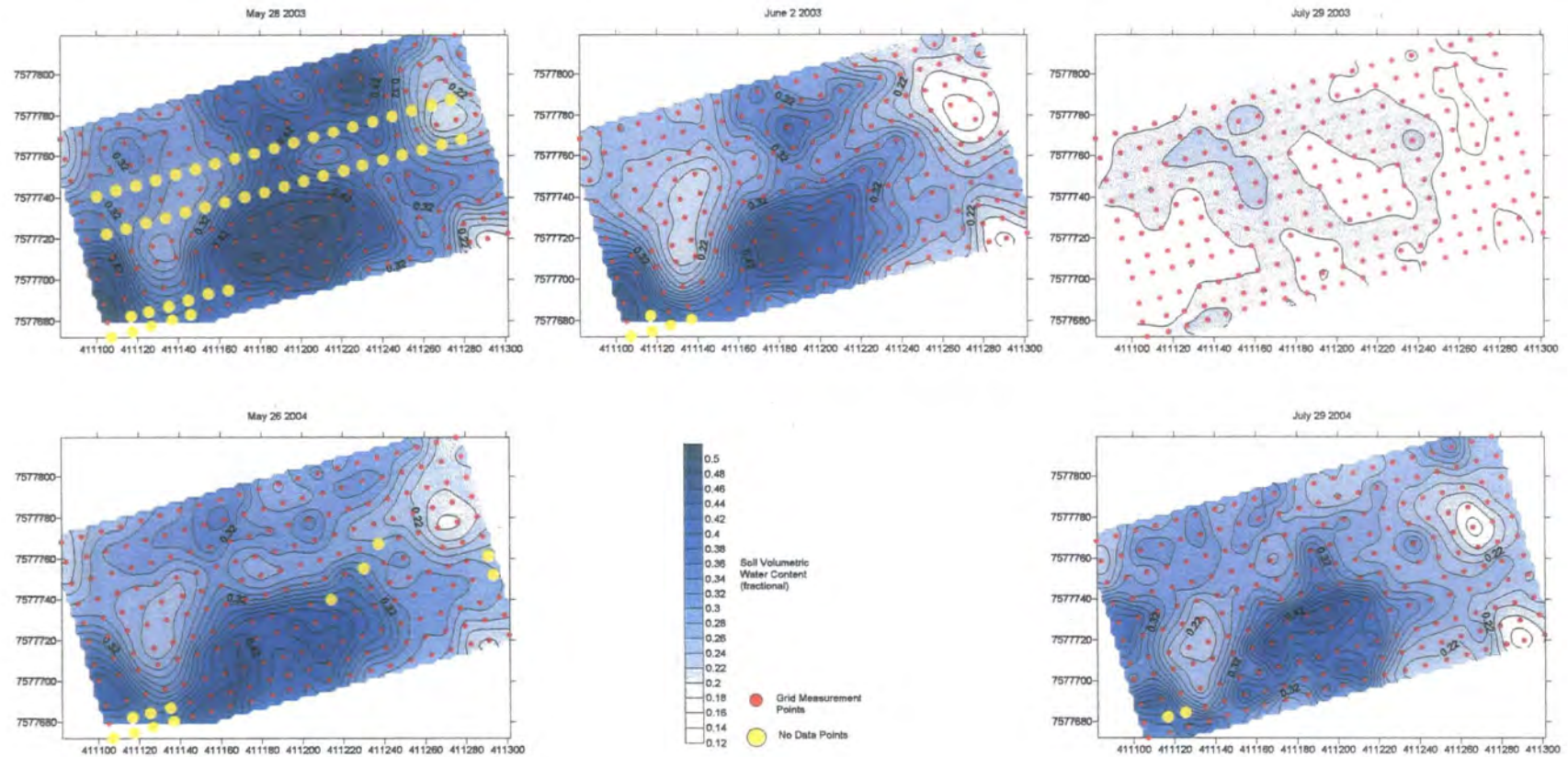
To aid data collection and analysis, the grid measurement points were numbered (Figure 3.10). This was useful when describing the spatial pattern of soil moisture across the grids.

Figure 3.11 shows volumetric water content surfaces for each of the spatial surveys of the Hillslope Grid. The red dots represent the measurement points and the yellow dots, points where measurements could not be taken due to the presence of snow or frozen soil. It should be noted that the snow or frozen soils occurred in the topographic hollows that give the hillslope its stepped appearance. In May of both years a large area of wet soil can be seen in the centre of the grid (around grid points E10 and E11). In 2004 this extends to the top of the hillslope (to the south around grid points D3 and D4) and there is spatial continuity of the moisture pattern in the north-south direction as suggested by the directional variogram. This wetter area persists in June 2003 and may have been caused by the lateral redistribution of surface water from persistent snow in the upslope topographic hollow. The grid is considerably drier in July 2003 when localized extremes of increased soil moisture are absent. However, this wet area is seen in July 2004 centred on grid points F13 and G13, having moved to the north-east.



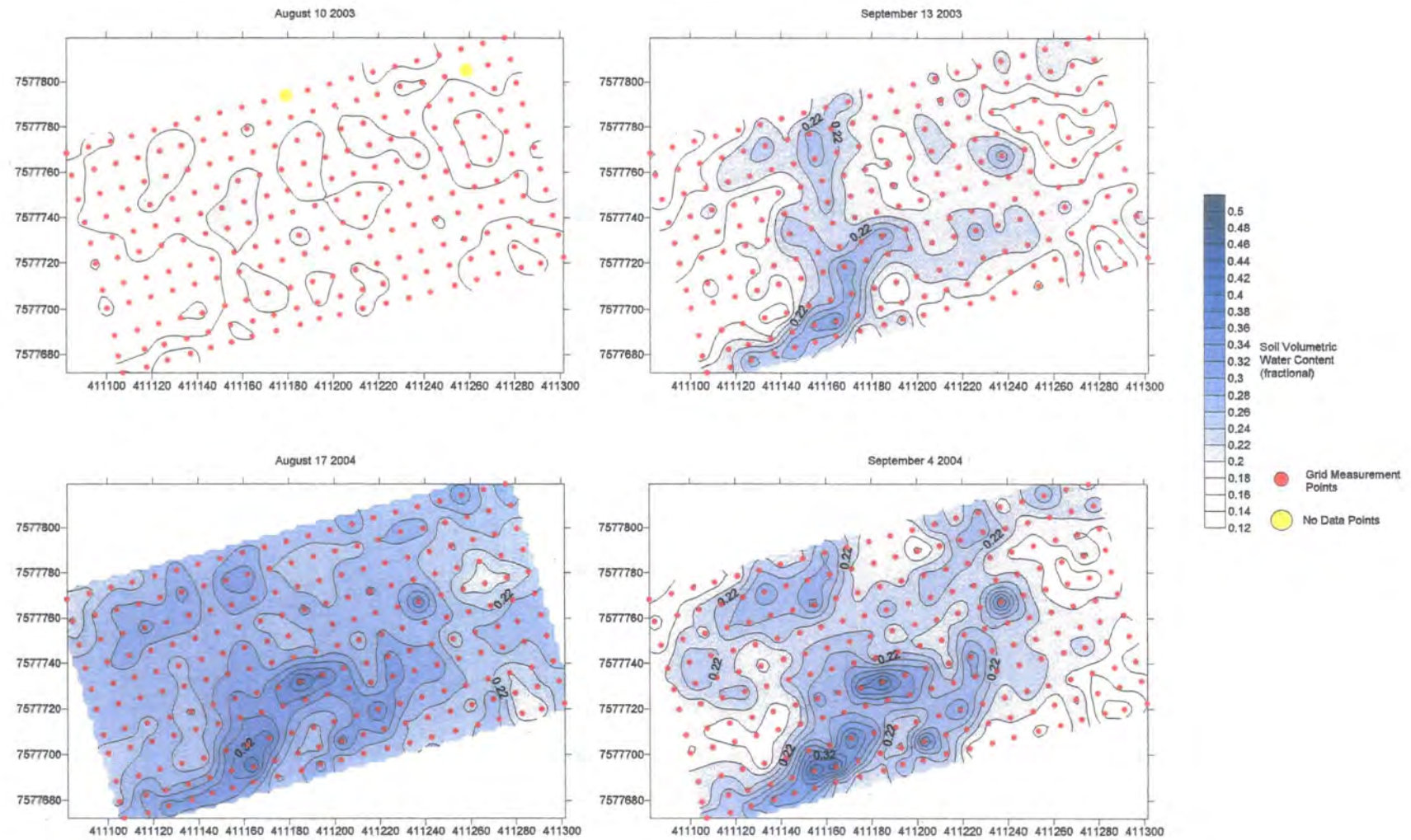


**Figure 3.11 Hillslope Soil Volumetric Water Content Surfaces Derived by Kriging using Spherical Variogram Models of the Spatial Soil Moisture Survey Data**



**Figure 3.12a Windfetch Soil Volumetric Water Content Surfaces Derived by Kriging using Exponential Variogram Models of the Spatial Soil Moisture Survey Data**





**Figure 3.12b Windfetch Soil Volumetric Water Content Surfaces Derived by Kriging using Exponential Variogram Models of the Spatial Soil Moisture Survey Data**

The snow in the south-west corner of the grid is not as extensive and the surface water ponding at point B'20 is present. As well as the south-west corner, points D'1 and F'7 were also snow covered and points E'1 and E'9 were frozen. Ponding was present at point G'6. No survey was carried out in June 2004 but in July the grid was much wetter than the year before. The pattern appears similar to that seen in June 2003, with a large wet area in the southern central and western parts of the grid. The two data points excluded in the south-west corner were waterlogged. The August plot shows an overall reduction in soil moisture but the wet areas in the southern half of the grid persist (centred on point B'15 spreading north-east into the centre of the grid). The drier areas shown in the 2003 plots (at A'2 and G'3) are also seen in this plot, along with the wet points at G'6 and K'13 and K'14. These features are seen in the September 2004 plot which appears wetter than in 2003.

#### 3.2.4 Time Series Soil Moisture and Rainfall

During the melt period, the main source of soil moisture is the melt water from spatially distributed snow cover. During this period, the presence of ground ice is critical when considering soil moisture distributions. Frozen ground, underneath spatially distributed snow cover, persists longer than in snow free areas. This means that melt water is typically redistributed, laterally as overland flow, to downslope areas of unfrozen ground. However, in the summer months it is rainfall events that provide the main input into the hydrological system. Using data collected by the automatic loggers, a fine resolution time series of soil moisture and rainfall was produced which allowed the soil moisture state to be traced throughout the study period.

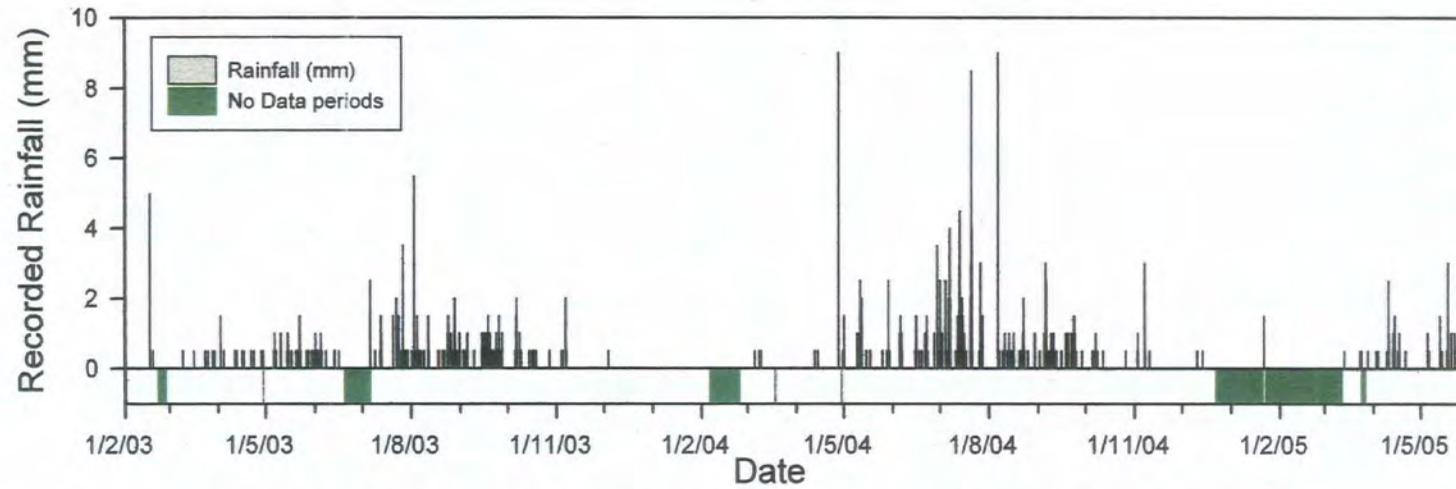
Figure 3.13 shows the recorded rainfall (half hourly averages) on the Windfetch Grid (from AWS1) during the study period. Periods of no available data (when technical difficulties prevented data collection) are marked on the graph to distinguish these periods from times of no rainfall. It should be noted that, during periods of snowfall and snow cover, the performance of the rain gauge is uncertain. During the summer of 2003, there were a number of small rainfall events (less than 2 mm) but only three events over



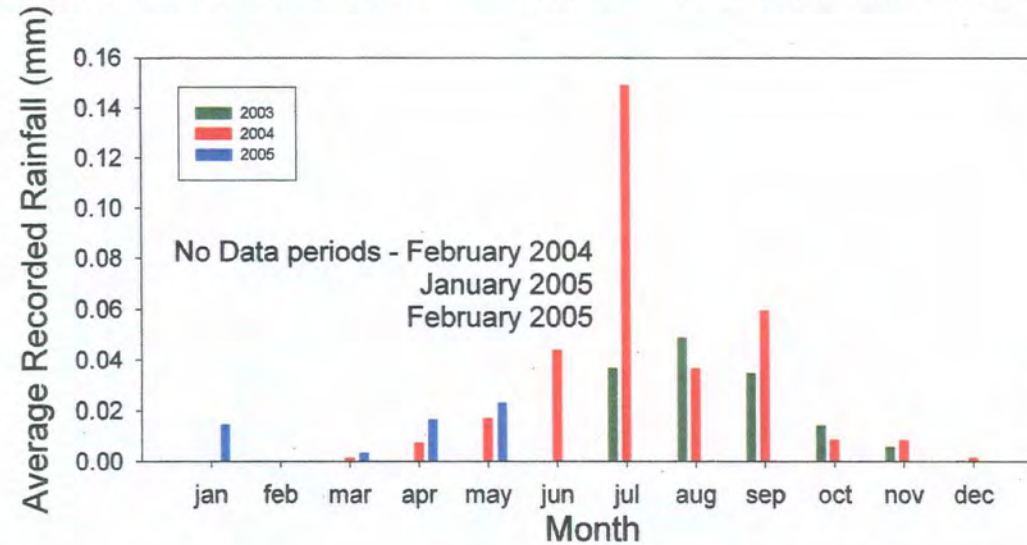
2 mm. In contrast, 2004 was a much wetter summer with many more rainfall events over 2 mm, the highest of these being 9 mm at the end of April and beginning of August. Figure 3.14 (monthly average rainfall) shows that July 2004 was by far the wettest month during the study period (the recorded rainfall running from 29<sup>th</sup> January 2003 to 31<sup>st</sup> May 2005).

The recorded rainfall data were then combined with the time series of soil moisture data recorded at the Hillslope and Windfetch sites to produce Figures 3.15 and 3.16 (respectively). At the *Empetrum* site on the hillslope (Figure 3.15), the soil water content at 5 cm is considerably less than at 10 cm during the summer of 2003. A number of wetting and drying curves are present at this time which corresponds to rainfall events. The soil moisture values drop during the soil freezing (in early November) due to technical difficulties with the CS616 sensors. In 2004, the recorded soil moisture values are higher than 2003, the wetting ‘peaks’ and drying ‘troughs’ being very steep in response to the numerous, large rainfall events. The *Salix* site shows a similar pattern of wetting and drying but the 5 cm soil moisture values are much closer to the 10 cm values (and are also higher) than at the *Empetrum* site. Also, during the rainfall events in 2004, the 5 cm soil moisture at the *Salix* site appears to remain at high values throughout the summer, the soil not drying down as much at this depth in comparison to the *Empetrum* site.

The soil moisture trace at the Windfetch site is shown in Figure 3.16. The period of frozen soil has been indicated on the graph. During this time, the very low soil moisture values observed until the beginning of May 2004 are due to technical difficulties with the sensors operating in frozen soils and are not true representations of the soil moisture state at this time. Once the soil had thawed (at the end of April at 5 cm depth), the soil moisture rose to high levels and remained very high for the duration of the summer (the gap in data was due to technical problems with the data collection). The soil moisture does not appear to go through the same wetting and drying cycles due to the rainfall events that are observed at the Hillslope site. It appears that, at this site, the soil remained at near-saturation levels throughout the summer of 2004.



**Figure 3.13 Recorded Rainfall on the Windfetch Grid for the duration of the study period**



**Figure 3.14 Monthly Average Recorded Rainfall (mm) on the Windfetch Grid during the study period (July 2003 – May 2005)**

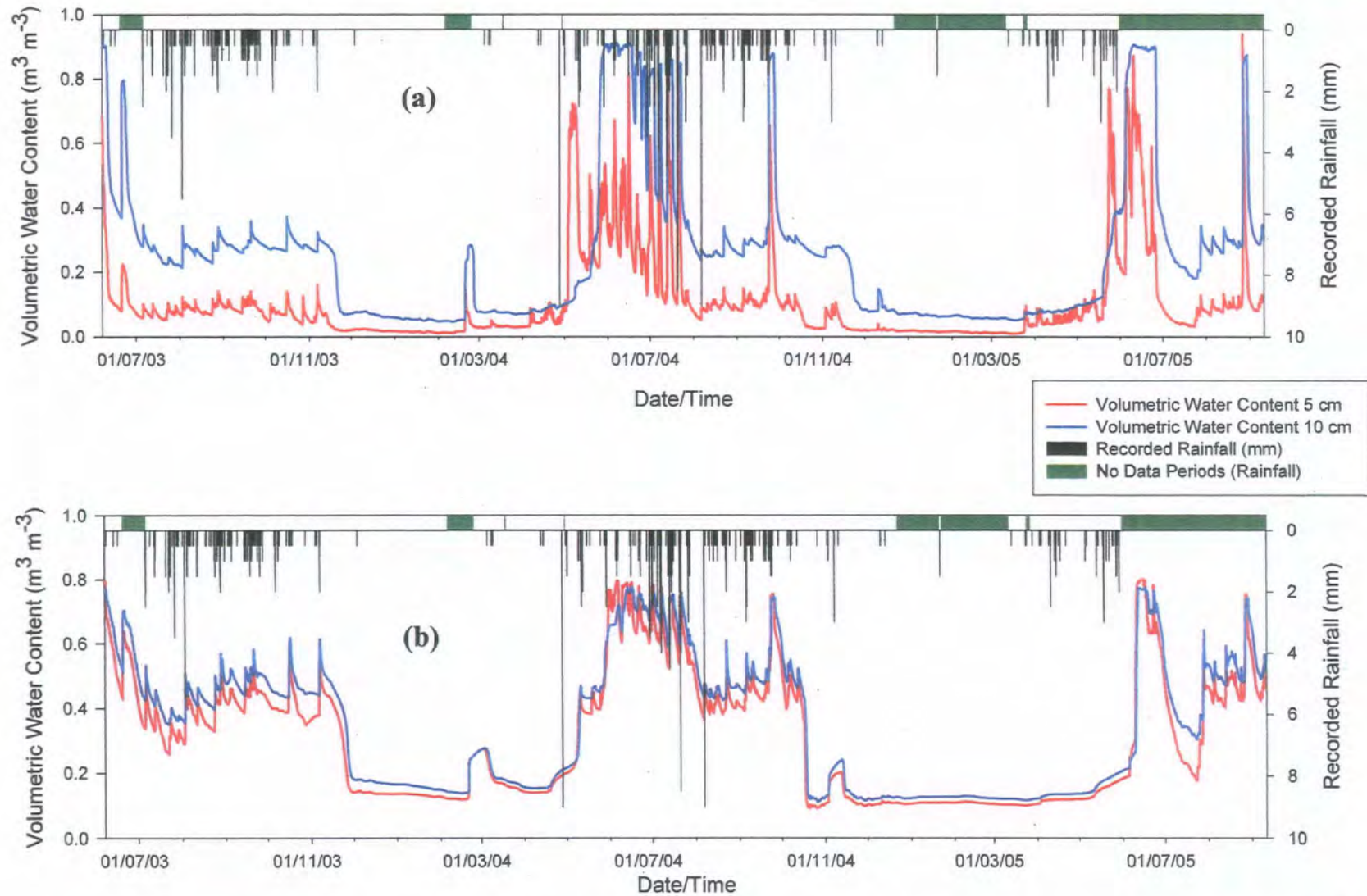
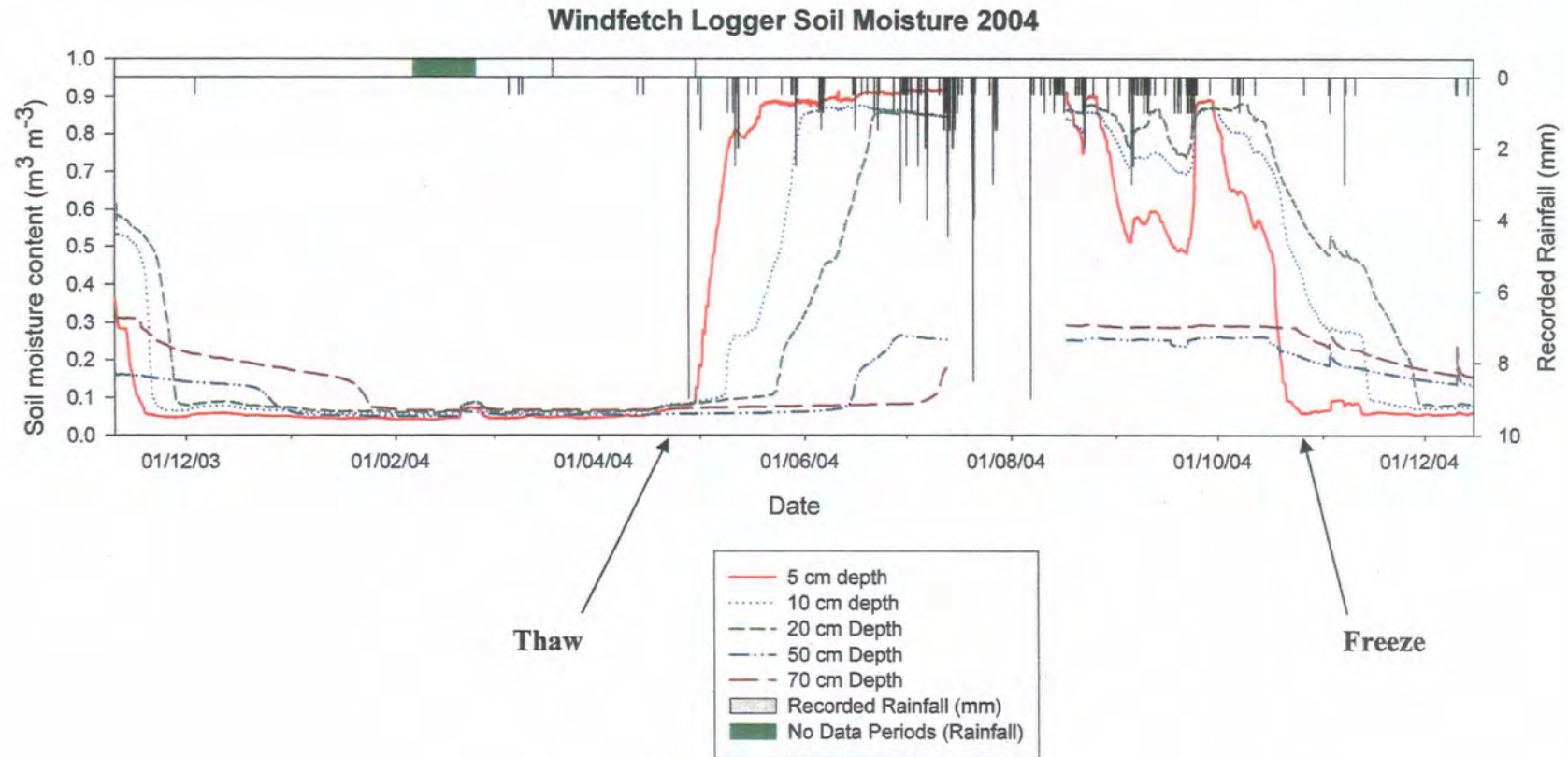


Figure 3.15 Hillslope Grid Volumetric Soil Water Content and Recorded Rainfall for the *Empetrum* (a) and *Salix* (b) sites during the study period



**Figure 3.16 Soil Volumetric Water Content at various soil depths and Recorded Rainfall for the Windfetch Grid**



### 3.2.5 Soil Freezing and Thawing

The spatial patterns of soil moisture from the SCIP surveys have provided a description of the distribution of volumetric water content in the top 5 cm of soil for both grids during 2003 and 2004. These surveys have shown that areas of both grids were still snow covered and frozen during the spring months of May and June. One problem with these surveys is that they do not give an indication of when the soil freezes and thaws since they are effectively a snapshot of the soil moisture state on the date of the survey. The automatic data collection stations, however, give a much better time series of soil moisture and soil temperature throughout the study period.

Figure 3.17 shows time series plots of volumetric water content and soil temperature for the *Empetrum* and *Salix* sites on the Hillslope Grid. The data were recorded from the 7<sup>th</sup> June 2003 to the 10<sup>th</sup> September 2005. During this period, the maximum soil temperature at 5 cm depth was 13.09 °C at the *Empetrum* site and 18.05 °C at the *Salix* site. The minimum temperatures recorded were -9.64 °C and -3.24 °C respectively.

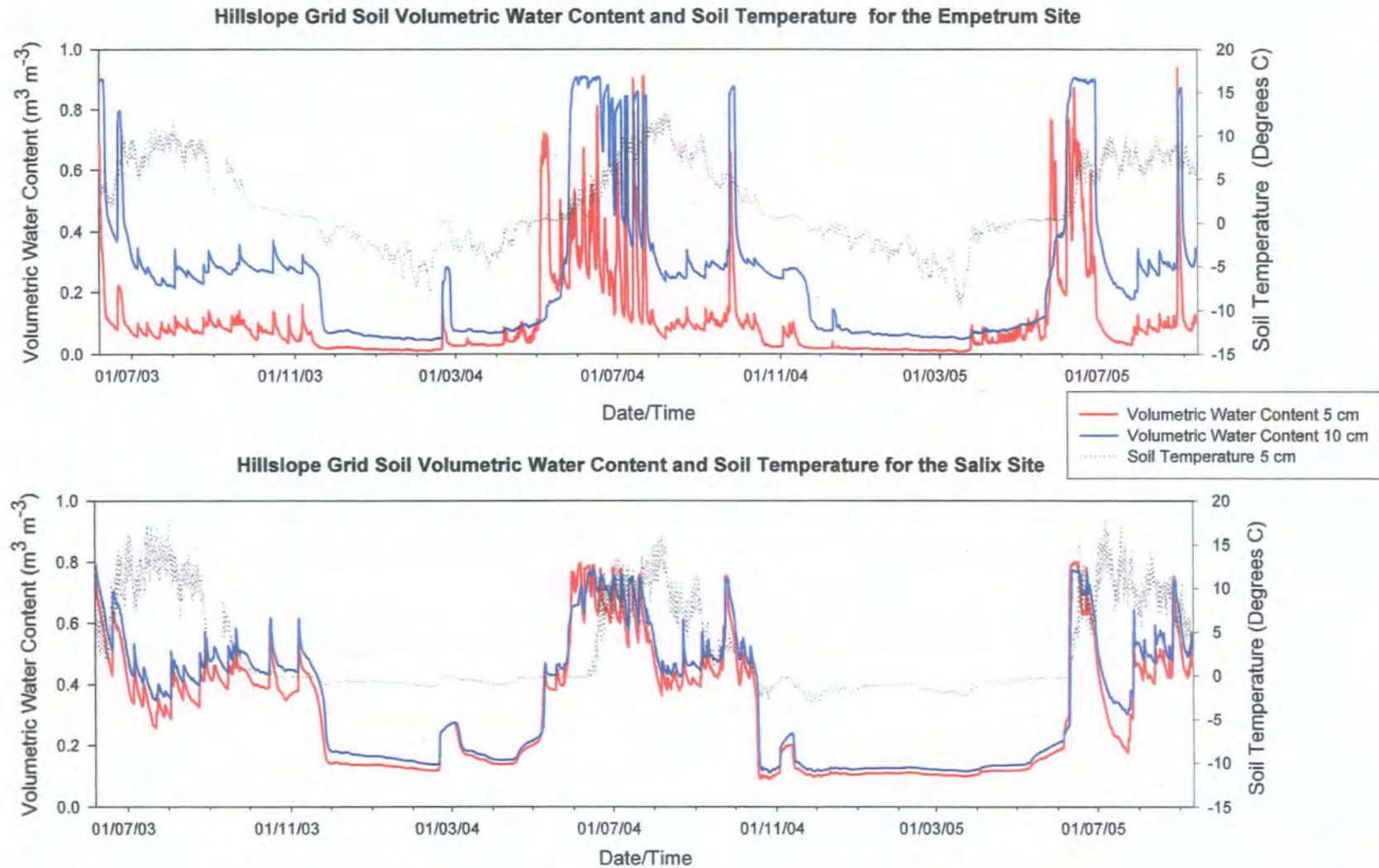
During the summer months of 2003, the volumetric water content at the *Empetrum* site showed a series of wetting peaks and drying troughs at both 5 cm and 10 cm depths (due to rainfall). The last of these wetting events occurred before the freeze on the 7<sup>th</sup> November. The soil temperature then dropped below zero on the 19<sup>th</sup> November and the mean daily soil temperature remained below zero for a number of days indicating the start of the freeze on this date. The soil moisture decreased rapidly at this time and became stable at a low soil moisture value for the duration of the freeze. In February 2004, the soil temperature rose above zero between the 20<sup>th</sup> and 25<sup>th</sup> (reaching a maximum of 0.23 °C) before falling to below zero again. During this period of elevated soil temperature, the soil moisture at both depths increased rapidly and a peak of 0.28 m<sup>3</sup>m<sup>-3</sup> was observed at a depth of 10 cm. This is comparable with the value of soil volumetric water content at this depth before the initial freeze event in November. The soil was then frozen until a thaw event between the 18<sup>th</sup> and 23<sup>rd</sup> April 2004. The soil temperature then dropped below zero until the 28<sup>th</sup> April when the soil thawed again and soil temperatures rose consistently during the spring months. During

the initial thaw event, the soil moisture at 5 cm depth increased slightly but no change was observed in the 10 cm readings. After the main thaw event, the soil water content remained at a low value until the 3<sup>rd</sup> May when the 5 cm readings rapidly increased from  $0.11 \text{ m}^3 \text{ m}^{-3}$  to around  $0.64 \text{ m}^3 \text{ m}^{-3}$  on the 4<sup>th</sup> May and up to  $0.72 \text{ m}^3 \text{ m}^{-3}$  on the 8<sup>th</sup> May. The 10 cm soil water content during this time showed only a slight rise. At this depth the water content increased initially between the 17<sup>th</sup> and 22<sup>nd</sup> May from  $0.18 \text{ m}^3 \text{ m}^{-3}$  to  $0.31 \text{ m}^3 \text{ m}^{-3}$ . The values recorded then show a large increase from  $0.31 \text{ m}^3 \text{ m}^{-3}$  to  $0.9 \text{ m}^3 \text{ m}^{-3}$  on the 30<sup>th</sup> May. The volumetric water content during the main thaw period was much higher than that recorded at the time of the freeze in November 2003. In 2004, the freeze began on the 20<sup>th</sup> November and the soil moisture values dropped to fairly constant values around this time (as in 2003). In December there were two periods when the soil temperature increased above zero for several hours. The first time this occurred, the 10 cm soil moisture increased by approximately  $0.06 \text{ m}^3 \text{ m}^{-3}$  but the 5 cm readings showed only a very small increase. The soil moisture values at both depths are not affected by the second occurrence of increased soil temperature in December. The thaw occurred in early April when the soil temperature rose above zero on the 2<sup>nd</sup> until the 6<sup>th</sup> before freezing again until the 16<sup>th</sup>. During this time there were only small changes in the soil water content at both depths (the 10 cm readings showing a slight but steady increase).

At the *Salix* site, the freeze occurred almost a month earlier than the *Empetrum* (22<sup>nd</sup> October) and the soil moisture continued to show cycles of wetting and drying up until the 28<sup>th</sup> November when the soil temperature reached  $-1^\circ\text{C}$  and the soil moisture dropped rapidly. The soil temperature increased around the 20<sup>th</sup> February (as in the *Empetrum* site data) but did not rise above zero. The soil moisture increased at this time before gradually dropping to levels slightly higher than before the event. The soil temperatures did not become consistently positive until the 14<sup>th</sup> June 2004. However, the soil moisture increased between the 8<sup>th</sup> and 10<sup>th</sup> May to approx  $0.46 \text{ m}^3 \text{ m}^{-3}$  at both depths. Some wetting and drying cycles are observed before the soil moisture increases to  $0.79 \text{ m}^3 \text{ m}^{-3}$  (5 cm) and  $0.71 \text{ m}^3 \text{ m}^{-3}$  (10 cm) around the 6<sup>th</sup> June. The 2004 freeze occurred on the 17<sup>th</sup> October and the soil moisture rapidly declined from  $0.34 \text{ m}^3 \text{ m}^{-3}$  (5 cm) and  $0.43 \text{ m}^3 \text{ m}^{-3}$  (10 cm) to  $0.09 \text{ m}^3 \text{ m}^{-3}$  (5 cm) and  $0.12 \text{ m}^3 \text{ m}^{-3}$  (10 cm) on the 20<sup>th</sup> October. The 2005 thaw occurred on the 9<sup>th</sup> June

and the soil moisture rapidly increased at this time to a maximum of  $0.8 \text{ m}^3 \text{ m}^{-3}$  (5 cm) and  $0.77 \text{ m}^3 \text{ m}^{-3}$  (10 cm) on the 15<sup>th</sup> June.

It should be noted that the soil temperature at the *Salix* site was considerably higher and more stable during the winter (although still below zero) than at the *Empetrum* site. This may be due to the greater snow depth at the *Salix* site (the site being in a topographic hollow) insulating the soil from the effects of extremes in air temperature. This will be investigated in a later chapter.

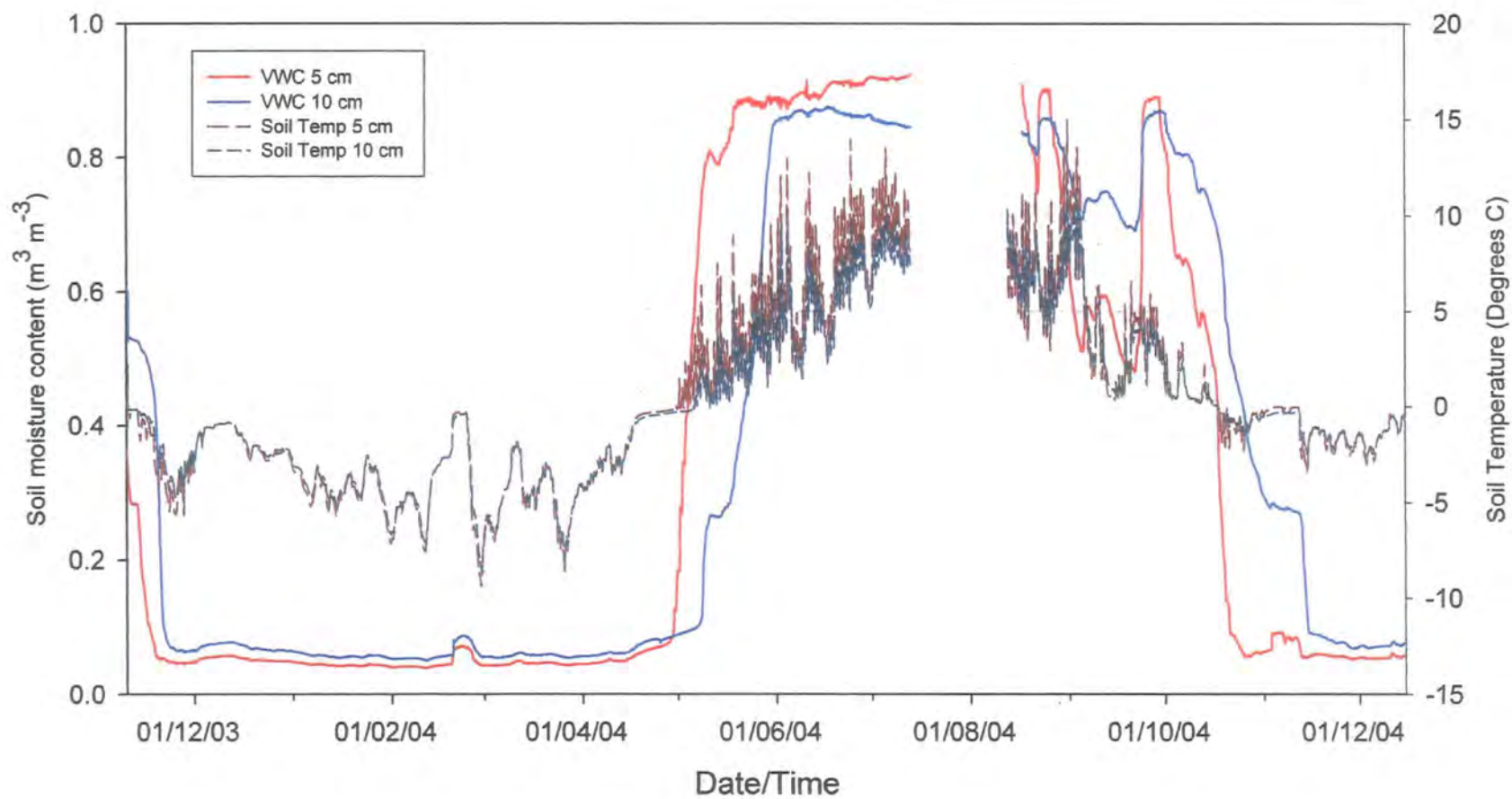


**Figure 3.17 Hillslope Grid Volumetric Water Content and Soil Temperature during the study period using CS616 Water Content Reflectometers and T107 Soil Temperature Probe**

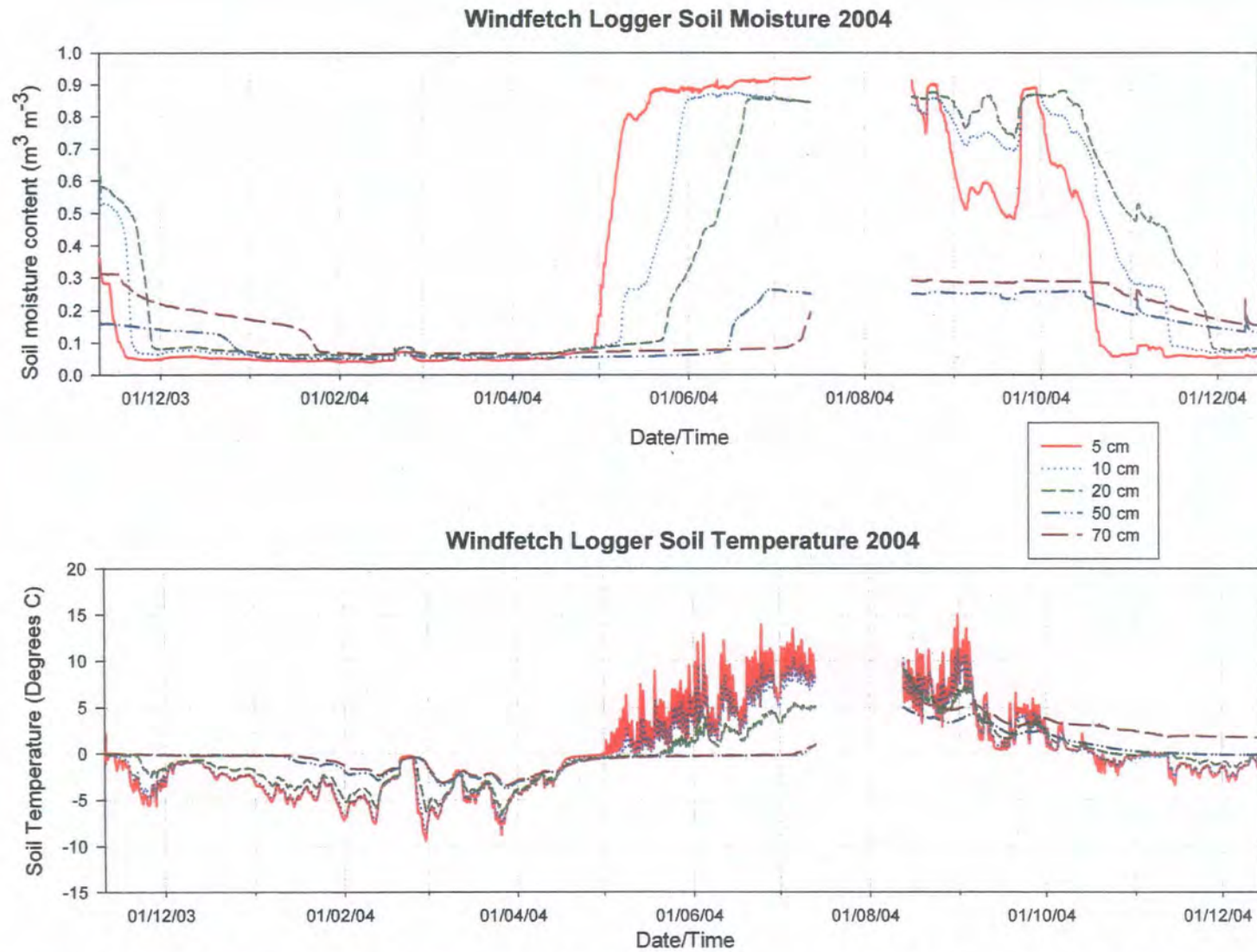


Figure 3.18 shows the volumetric water content and soil temperature at 5 cm and 10 cm depths at the Windfetch logger site. At this site the soil temperature rose above zero briefly on the 30<sup>th</sup> April 2004 at 5 cm depth and was consistently thawed on the 2<sup>nd</sup> May. At this soil depth, the soil moisture increased from  $0.16 \text{ m}^3\text{m}^{-3}$  (30<sup>th</sup> April) to  $0.53 \text{ m}^3\text{m}^{-3}$  (5<sup>th</sup> May) and continued to rise throughout May. At 10 cm depth, the soil temperature became positive on the 5<sup>th</sup> May for a few hours and then consistently thawed on the 6<sup>th</sup> May. The soil moisture started to gradually increase at this time before rising more rapidly between the 7<sup>th</sup> and 12<sup>th</sup> May. It took approximately two weeks for the soil moisture at 10 cm depth to rise to a maximum of  $0.86 \text{ m}^3\text{m}^{-3}$  (4<sup>th</sup> June). The soil freezing occurred at both 5 cm and 10 cm depth at approximately the same time (17<sup>th</sup> October). The 5 cm soil water content at this time reduced steadily but the 10 cm soil water content was a lot slower to respond to the soil freezing. At 5 cm depth, the soil moisture reduced from  $0.45 \text{ m}^3\text{m}^{-3}$  on the 17<sup>th</sup> to  $0.25 \text{ m}^3\text{m}^{-3}$  on the 19<sup>th</sup>. During the same period, the 10 cm values changed from  $0.69 \text{ m}^3\text{m}^{-3}$  to  $0.67 \text{ m}^3\text{m}^{-3}$ .

The Windfetch datalogger site recorded soil volumetric water content and soil temperature at 5 cm, 10 cm, 20 cm, 50 cm and 70 cm soil depth. A time series plot of these data can be seen in Figure 3.19. The volumetric water content at 5 cm, 10 cm and 20 cm depths were seen to rise during May as the soil temperature increased. There appears to be a lag between the 5 cm and 10 cm and again between the 10 cm and 20 cm soil moisture content rising to their maximum values. The soil temperature plots show that at 5 cm, 10 cm and 20 cm depths, the temperatures are much more variable than at deeper depths. The 50 cm and 70 cm temperatures remained at around zero for much longer than the shallower soil temperatures.



**Figure 3.18 Soil Volumetric Water Content and Soil Temperature at 5 cm and 10 cm Depth for the Windfetch Grid**



**Figure 3.19 Soil Volumetric Water Content and Soil Temperature at various depths for the Windfetch Grid**

### 3.3 Chapter Summary

Field measurements were undertaken to provide a spatial and temporal description of the soil moisture and temperature regime. Manual near surface soil moisture measurements were made across two contrasting grids (the sloping Hillslope Grid and the flatter Windfetch Grid) over two summer seasons. These were complemented by fine resolution, continuous time series data collected automatically at a number of locations (and soil depths) for the duration of the study period. These data were then used to evaluate the processes controlling soil moisture variability across the study area.

The highest soil moistures were recorded during the spring months (May and June) of both 2003 and 2004 across both sites. This is due to melting snow which is the main water input into the hydrological system at this time and persist into the early summer months. Histograms demonstrate that seasonality is displayed in the near surface soil moisture values for the two sites (Figures 3.2 and 3.3).

Omnidirectional variograms produced for both the Hillslope and Windfetch sites (Figures 3.5 and 3.6) show high sills during the spring months (May and June) and progressively lower sills with time through the summer months. These findings suggest that the seasonal differences in sill values may be due to the different processes controlling soil moisture in wet and dry conditions. Redistribution of water due to topographic factors during wet periods causes high sills whereas, during drier conditions, soil moisture is relatively consistent (being controlled by factors such as soil properties and evapotranspiration) leading to lower sill values. The range (correlation lengths) determined for this study were between 50 m and 55 m. These are similar to correlation lengths described in the literature carried out at a similar scale (Nyberg, 1996, Western *et al.*, 1998a, Anctil *et al.*, 2002). The variability of nugget values from the variogram modelling (higher nuggets during wetter periods) is also consistent with the findings of other studies (Grayson *et al.*, 1997; Western *et al.*, 1998a; 1998b; 1999a; 2003; Wilson *et al.*, 2004).

Directional variograms (Figures 3.7 and 3.8) show that during the snowmelt period, the Hillslope site displays a higher variability in the east-west direction than in the north-south direction. These findings are significantly different to the omnidirectional variogram, suggesting that the structure is anisotropic. Topography therefore plays an important part in the soil moisture patterns observed. Later in the season, the directional variograms show little variation and are considered to be isotropic. This shows that topography is not the main driving factor for the soil moisture patterns observed during the drier periods.

Spatial soil moisture distributions, derived from the manual surveys, suggest that two soil moisture states can be distinguished (wet and dry) and that switching occurs between these two states indicating whether the process controlling soil moisture patterns is caused by lateral or vertical water fluxes. During the snowmelt period, the soil is saturated and moisture patterns are controlled by the lateral movement of water (caused by topography). In the dry state (July, August and September 2003), the moisture patterns are determined by local influences such as soil properties and evapotranspiration. As the soil dries, soil water suction increases and soil hydraulic conductivity decreases. The effects of rainfall and evapotranspiration are then the main cause of the spatial soil moisture patterns.

Soil moisture time series data shows the rapid response of soil moisture across the site to rainfall events. On the Hillslope site, the *Salix* soils are wetter than the *Empetrum* soils which is due, in part, to the lateral movement of water since the site is located at the base of a topographic depression. Also, differences in soil depth, soil structure and soil organic content would account for the differences. Soil temperature time series data show distinct differences in the soil temperature regime between the *Empetrum* and *Salix* sites during freeze and thaw periods. When the soils are frozen, the soil temperature at the *Empetrum* site drops well below zero (to a minimum of approximately  $-9^{\circ}\text{C}$ ) and is seen to fluctuate greatly. In contrast, the soil temperature at the *Salix* site is noticeably higher (a minimum of approximately  $-3^{\circ}\text{C}$  was observed) and more stable during the winter months. This may be due to the insulating effect of the snow at the *Salix* site during this period, since the site is situated in a topographic hollow that accumulates wind blown snow.

## **Chapter 4**

**To assess the role topography plays in determining soil moisture distributions at the hillslope scale**

The Hillslope and Windfetch Grids were chosen for this study since they display distinctly different topographic features; the Hillslope Grid has areas of high slope angle giving it a 'stepped' appearance while the Windfetch Grid has much shallower slopes and is flatter. The hypothesis that the wet areas receive water from an upstream 'catchment' was developed to investigate how the wetter areas described in Chapter 3 received their water.

The upstream 'catchment' hypothesis suggests water is redistributed on the surface across the landscape by topographic factors such as slope angle and upstream contributing area. The wet areas observed across the site could also be due to a sub-surface catchment which intersects the surface in a spatially heterogeneous way.

This chapter will describe the topography of both the Hillslope and Windfetch Grids and will go on to investigate the degree to which the topography influences the spatial distribution of near surface soil moisture.

#### 4.1 Field Methods and Data Collection

To investigate the topographic influence on the spatial soil moisture distribution it was necessary to collect survey data so that a digital representation of the landscape could be produced. The two methods of data collection employed to collect these topographic data were Real Time Kinematic (RTK) surveying using differential GPS (Trimble Navigation Ltd) and remotely sensed LiDAR (Natural Environment Research Council Airborne Remote Sensing Facility). The RTK survey of the Hillslope Grid was undertaken in August and October 2004. This involved the collection of over 11,000 point measurements of location (Easting and Northing) and elevation of the land surface at one metre intervals and with centimetre accuracy across the hillslope. This method of land surface surveying proved to be very fieldwork intensive and, although valuable spatial data were collected, the RTK surveying method could not be conducted for the Windfetch Grid at the one metre scale. The topographic data for the Windfetch Grid was provided by an airborne remote sensing flight which used LiDAR to collect point measurements of location and elevation at

1.5 m resolution and centimetre accuracy. These data were collected on the 17<sup>th</sup> July 2005. Further details of the topographic surveying methods can be found in Section 2.4.

Spatial surveys using the SCIP provided the data for the production of the spatial patterns of soil moisture (outlined in Section 2.2.3 and described in detail in the previous chapter). The hillslope and Windfetch soil station data (described in Chapter 3) were used to provide soil moisture and temperature data during the study period.

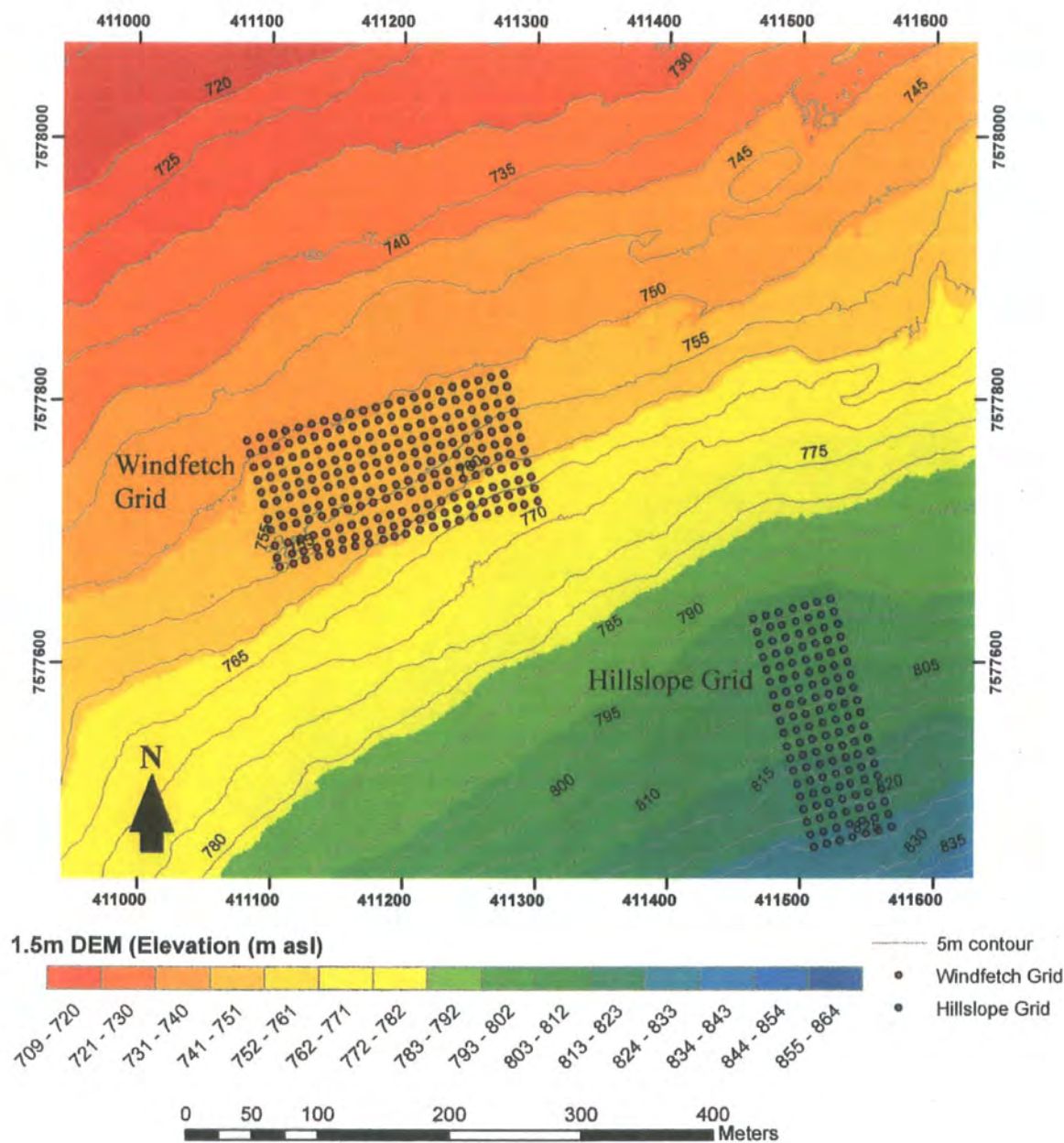
#### 4.2 Analysis of Data

In order to analyse the topographic influence on the spatial distribution of soil moisture across the study area, it was necessary to create a surface with a raster structure which involved some interpolation. This was achieved using the LiDAR dataset which covered the entire fieldsite at 1.5m resolution.

The LiDAR data were used to create a hydrologically correct Digital Elevation Model (DEM) of the study area using the TOPOGRID command in ArcInfo. This command uses an interpolation method (iterative finite difference interpolation) that has been designed specifically for hydrological purposes to take into account abrupt changes in terrain (ESRI, 2004).

Figure 4.1 shows the 1.5 m Digital Elevation Model of the STEPPS site and the location of the Hillslope and Windfetch Grid data collection points. This figure highlights the relative size and orientation of the grids on the landscape and their relative position. A 1.5 m DEM of the Windfetch Grid was then 'clipped' from the site DEM. The RTK survey data of the Hillslope Grid was used to create a 1 m resolution DEM (using the TOPOGRID command in ArcInfo) for the Hillslope site.





Projected Coordinate System:  
WGS 1984 UTM zone 34N  
Datum: WGS 1984

Figure 4.1 Location of the Hillslope and Windfetch Grids at the STEPPS fieldsite (displayed in ArcMap)

To create the hillslope DEM, the data from the RTK survey of the Hillslope Grid were first imported into the Trimble Geomatics Office software package for processing (Trimble Navigation Ltd.). This software was used to check for any data errors and to verify the quality of the data collected by the roving survey unit. The points collected by the hillslope topographic survey and the resulting point report can be seen in Figure 4.2.

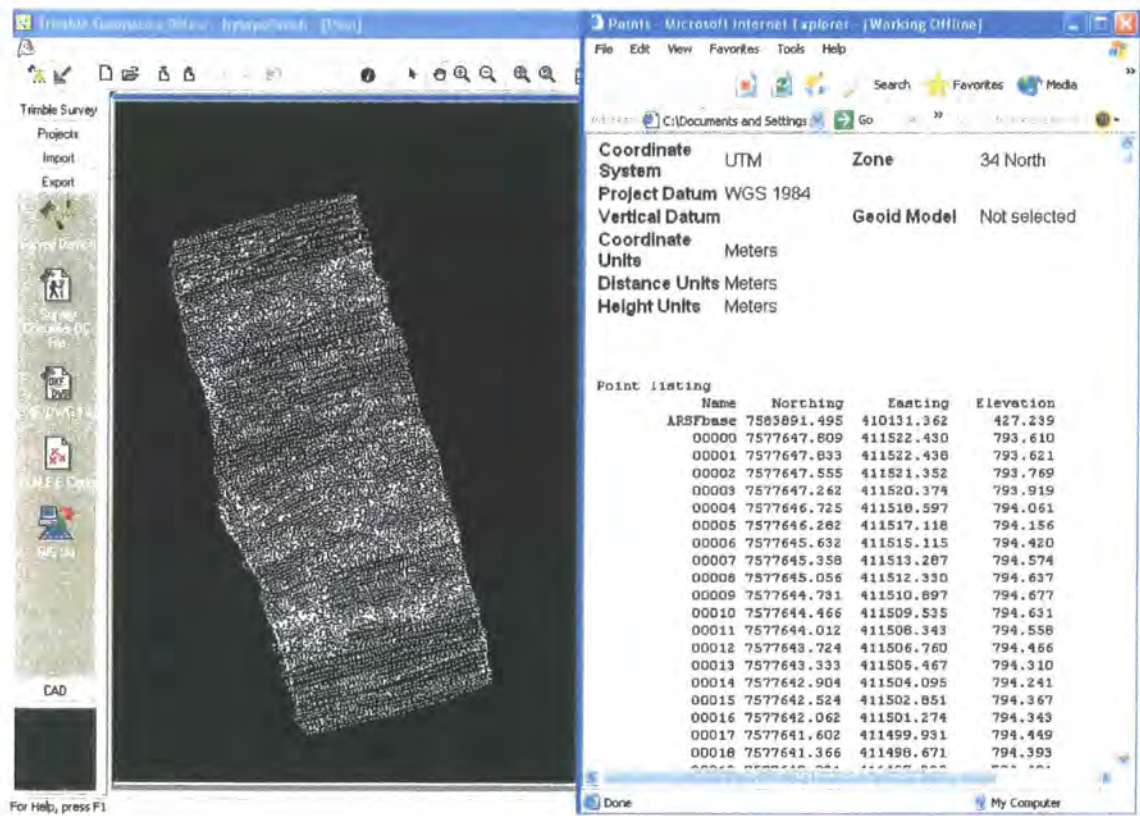


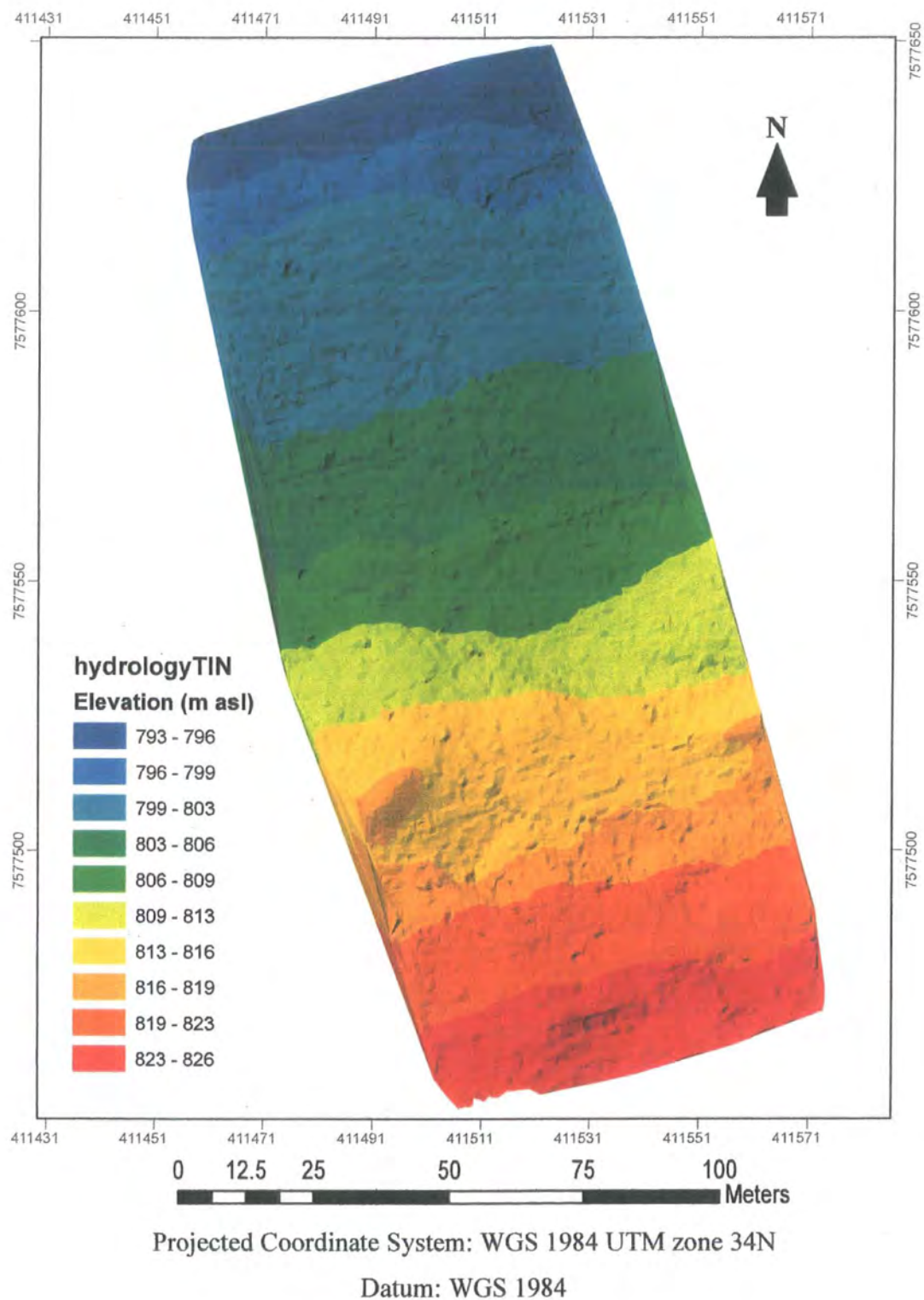
Figure 4.2 Trimble Geomatics Office software showing the data points collected on the Hillslope Grid and the format of the point data

The point data were then exported in an x,y,z ASCII text format to facilitate import and processing in ArcGIS 9 (ESRI, 2004). In order to represent the topography digitally, a Triangular Irregular Network (TIN) was produced using ArcInfo. The TIN method joined all the surveyed data points with their nearest neighbours to produce Delaunay triangles (where a circle can be drawn through the three nodes of a triangle that contained no other

data points) (ESRI, 2004). This produced a surface that could then be displayed in ArcMap and is shown in Figure 4.3.

The TIN allowed a representation of the hillslope to be produced which was based only on the data points collected and did not involve any spatial interpolation. Also, the TIN preserved the spatial precision of the original survey data and produced a high resolution surface in areas of high survey point density.





**Figure 4.3 Triangular Irregular Network (TIN) of RTK survey points for the Hillslope  
Grid displayed in ArcGIS**

Both the Hillslope and Windfetch Grids have an aspect in a north to north-westerly direction. The Hillslope Grid has a maximum elevation at grid point A1 (825.17 m a.s.l.) (grid point naming conventions used are those shown in Figure 3.10) and minimum at grid point G19 (793.68 m a.s.l.), a difference of around 32m from the bottom of the grid to the top. The Windfetch Grid has a maximum elevation at grid point A'1 (768.27 m a.s.l.) and a minimum at grid point L'21 (749.45 m a.s.l.), a difference over the grid of 18.82 m.

Slope maps were then derived from the LiDAR 1.5 m DEM (Windfetch) and 1 m RTK DEM (hillslope) using the 3D Analyst extension in ArcGIS (shown in Figure 4.4 below). Figure 4.4 shows the slope angle of the topography on the Windfetch (left) and Hillslope (right) Grids.

The circular areas of high slope on the Windfetch Grid (on the western edge and central northern area) are the result of anomalies (e.g. the eddy correlation tower) in the LiDAR data. Two linear slope features can be distinguished running in a north easterly direction across this grid (with a maximum slope of 35°) but slope angles are predominantly between 0.02° and 15°.

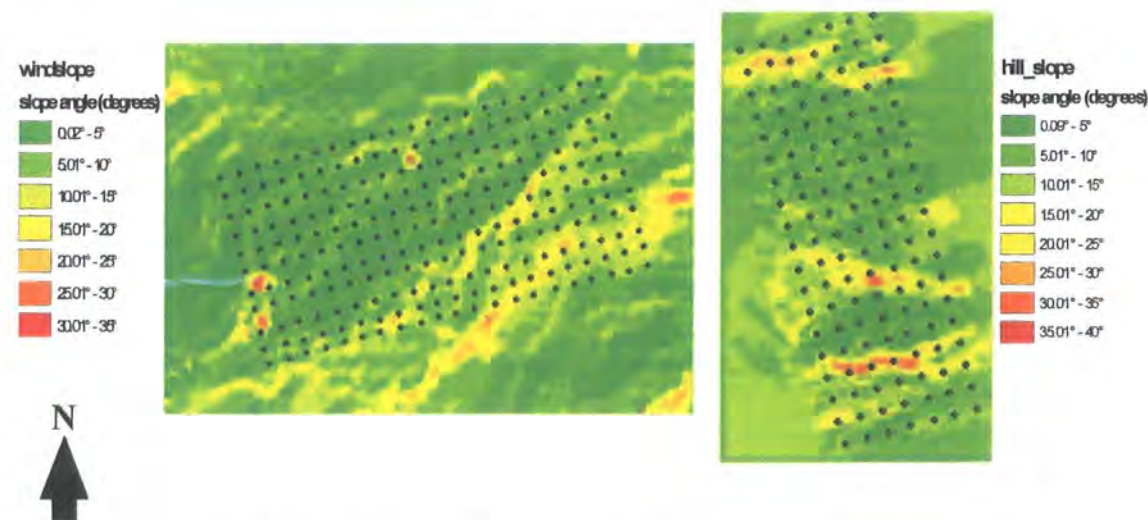
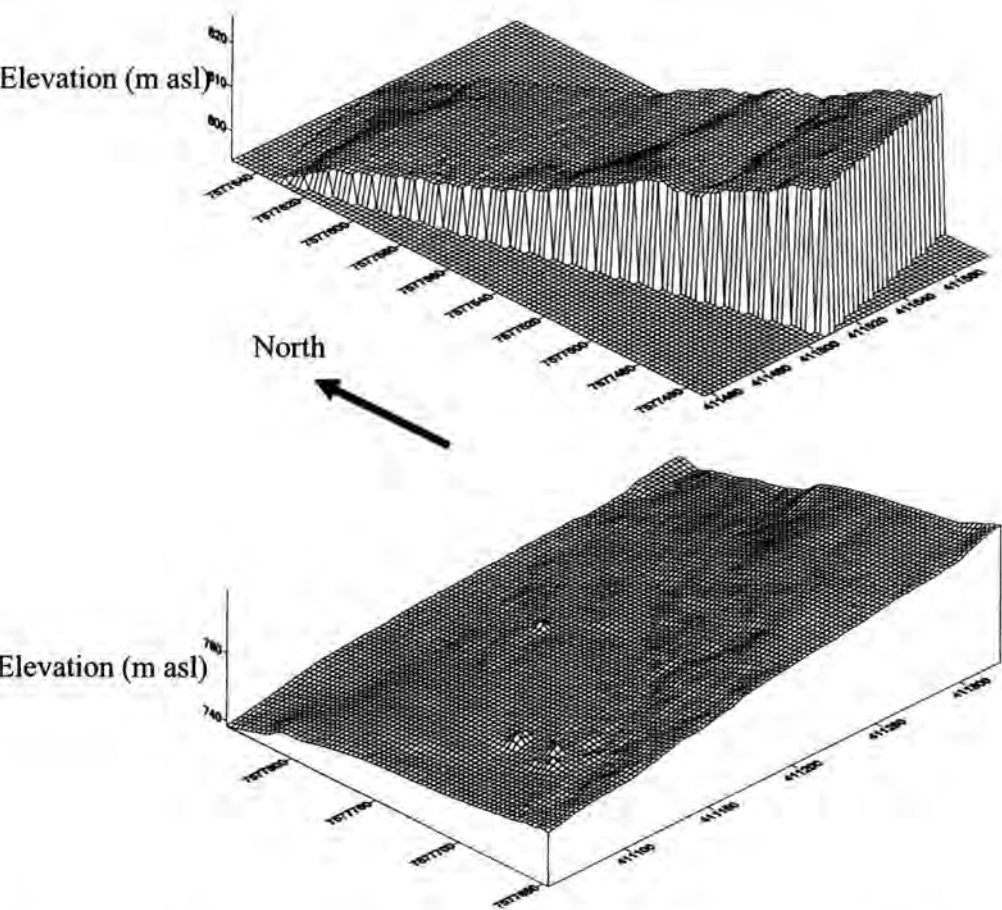


Figure 4.4 Slope maps of the Windfetch and Hillslope Grids

The Hillslope Grid has three areas of high slope angle running approximately east-west in direction. These slopes have a maximum slope angle of 40° and give the Hillslope Grid a distinctly ‘stepped’ appearance. A large area of flatter topography can be distinguished in the northern half of the grid.

In order to investigate the topographic influence on the spatial soil moisture patterns described in Chapter 3 (Figures 3.11 and 3.12), the topography was represented as a wireframe surface in the Surfer 7 software. The surfaces, shown in Figure 4.5, were produced using the RTK survey and LiDAR data using an ordinary kriging algorithm and linear model.



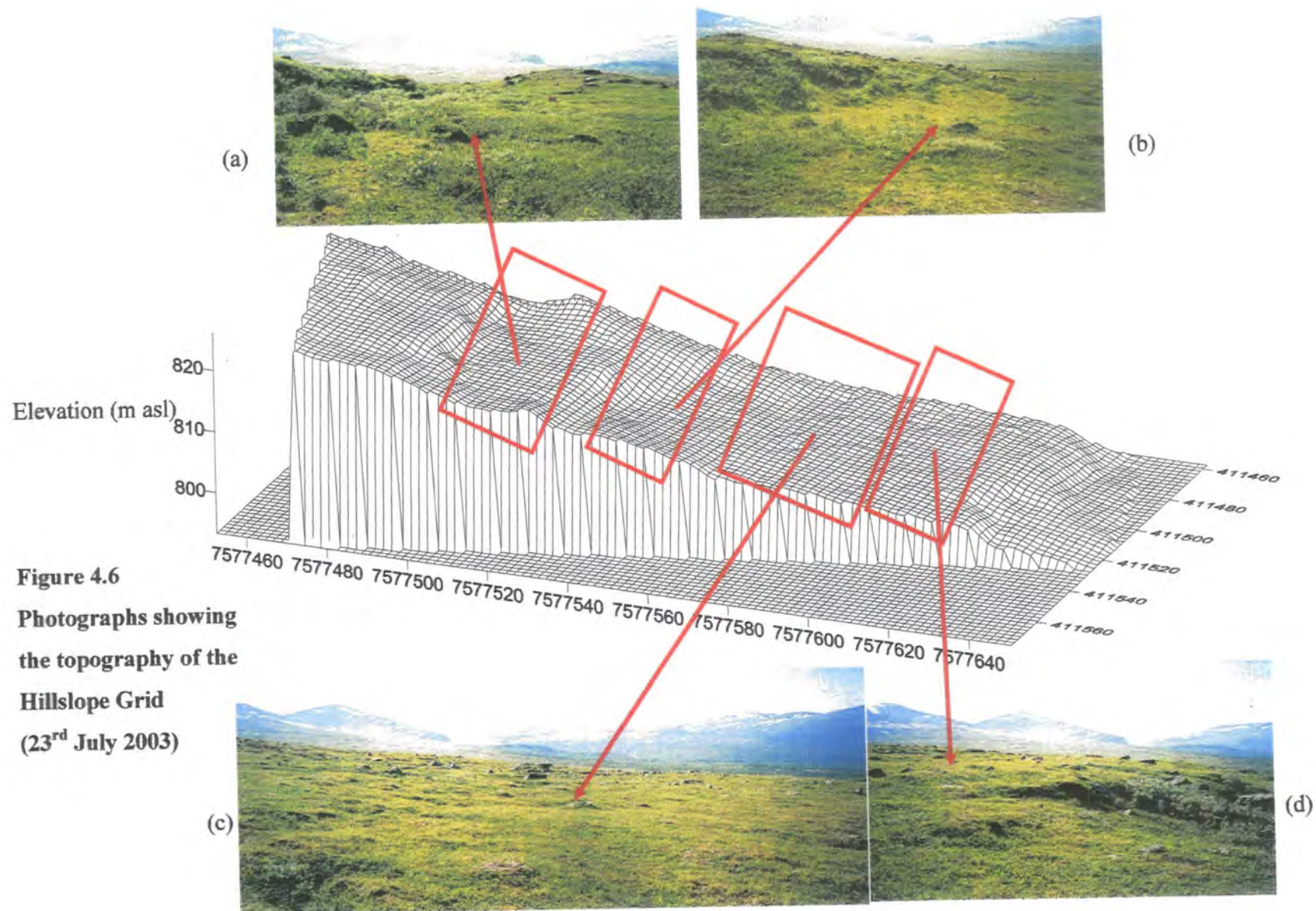
**Figure 4.5 Surfer 7 wireframe grids of the Hillslope (top) and Windfetch (bottom) sites**

Photographs of the Hillslope and Windfetch Grids are shown in Figures 4.6 and 4.7 respectively. The Hillslope Grid logger was placed at the base of the first (southern-most) area of high slope angle (topographic hollow) which can be seen in Figure 4.6 (a). The ‘stepped’ nature of the Hillslope Grid can be seen in Figure 4.6, the first topographic hollow leading on to a flatter plateau (a) before another area of high slope angle (b). The large area of flatter topography in the northern half of the grid can be seen in Figure 4.6 (c). This leads to the third ‘step’ at the base of the grid, the top of which can be observed in Figure 4.6 (d).

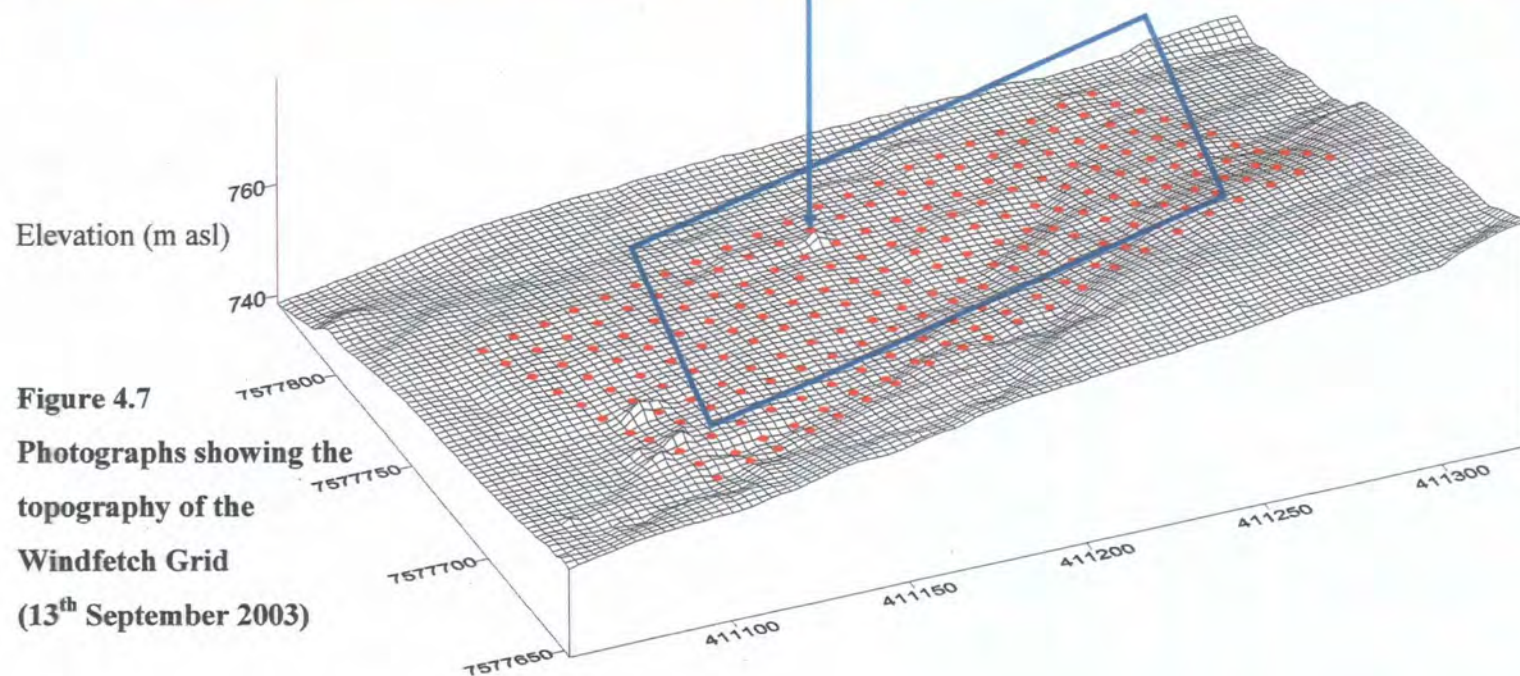
One of the most striking features of the hillslope observed in figure 4.6 is the variation in vegetation type associated with the topography. Areas of high slope angle are predominantly covered with *Salix* (Dwarf Willow) shrub which gives way to *Empetrum* Heath on the flatter areas. This is mainly due to the shelter offered by the topographic hollows from the effects of high winds which are observed at the site. The relationships between topography and land cover type will be discussed in more detail in Chapter 5.

Figure 4.7 shows a composite photograph of the Windfetch Grid. The Eddy Covariance tower and AWS1 can be seen (in the distance) in the centre of the photograph. An area of high slope angle can be made out in the foreground of the photograph. The topography of the Windfetch Grid is observed to be much flatter than that of the Hillslope Grid and the ‘stepped’ topographic features of the Hillslope Grid are not observed.





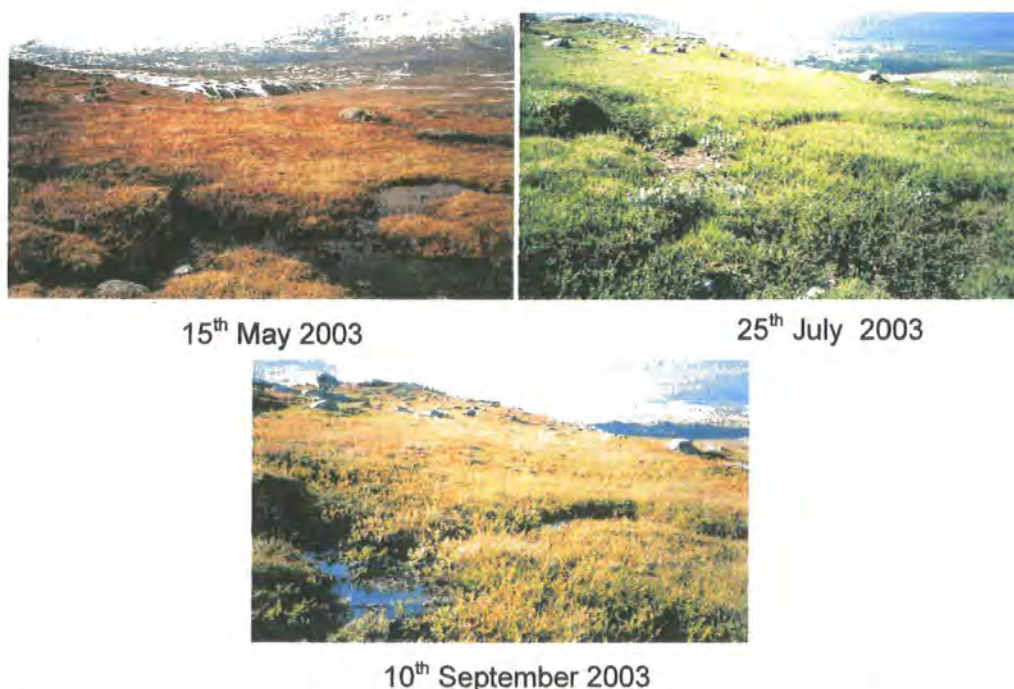




#### 4.2.1 Topography and Soil Moisture Distributions

The spatial distribution of soil moisture across both grids, described in Chapter 3, highlighted the locations of elevated soil wetness. In order to investigate the source of this soil moisture and why these wet areas persistently occur in specific locations it is necessary to use the topographic descriptions outlined above.

The topographic features described so far in this chapter are observed at the landscape scale (tens to hundreds of metres). It should be noted that plot scale (one to ten metres) topographic features are also present across the study area. Observations made in the field identified a network of channels and depressions (at a sub-metre scale) across both grids that were prone to flowing and standing water respectively. On the Hillslope Grid, a network of channels carried water into the grid from upslope during spring snowmelt. Figure 4.8 shows one of these channels to the south of the Hillslope Grid during the spring snowmelt (15<sup>th</sup> May), summer dry period (25<sup>th</sup> July) and approaching winter (10<sup>th</sup> September).



**Figure 4.8 Photographs of a ‘stream’ channel to the south (upslope) of the Hillslope Grid**

The channel shown in Figure 4.8 appears to be the main source of surface water flow into the Hillslope Grid during wet periods. When the channels are traced through the grid they appear to end before areas of high slope angle (the topographic ‘dips’) where the water flows under the rocky face of the ‘dip’ before reappearing at the base as surface water.

These ‘stream’ channels were not surveyed in detail (using the Real Time Kinematic surveying method) but it is important to acknowledge the presence of such surface features when considering the redistribution of water over both grids (especially the Hillslope Grid with higher slope angles). These channels also suggest that water is being redistributed from upslope areas and gives credence to the ‘upstream catchment’ hypothesis.

Since the ‘upstream catchment’ hypothesis is topographically driven, it is useful to show the soil moisture distributions in relation to the topography of the grids so that the relationship between the two can be explored.

Figures 4.9 (a-d) and 4.10 (a-e) show the soil volumetric water content surfaces derived from the SCIP surveys (described in Chapter 3) mapped onto the three dimensional wireframe topographic surfaces for the Hillslope and Windfetch Grids respectively. Also included in these figures is the recorded rainfall (half hour averages) for the 7 days preceding the survey to give an indication of the water input due to precipitation.



#### 4.2.1.1 The Hillslope Grid

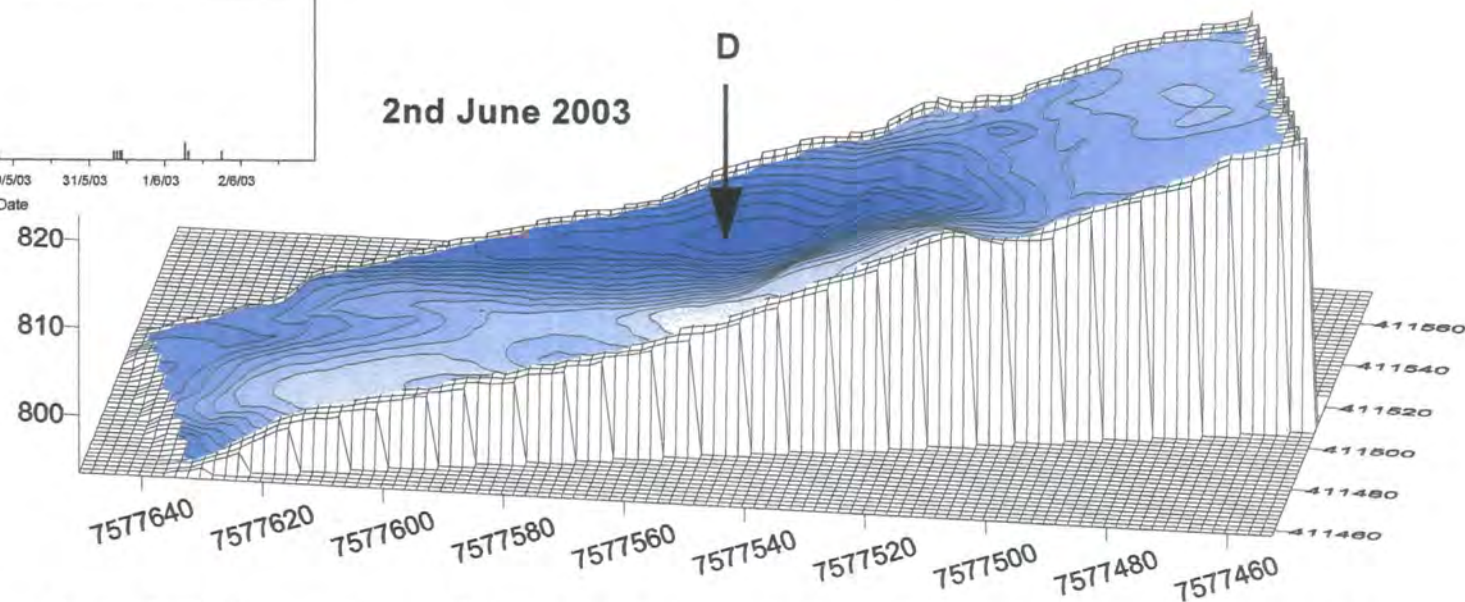
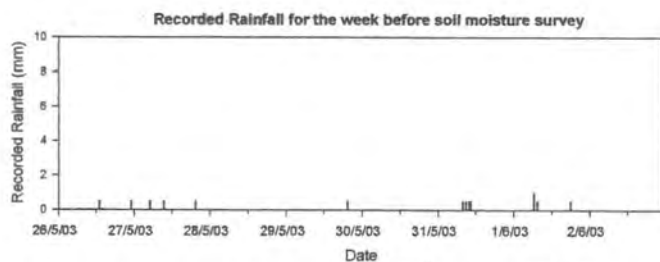
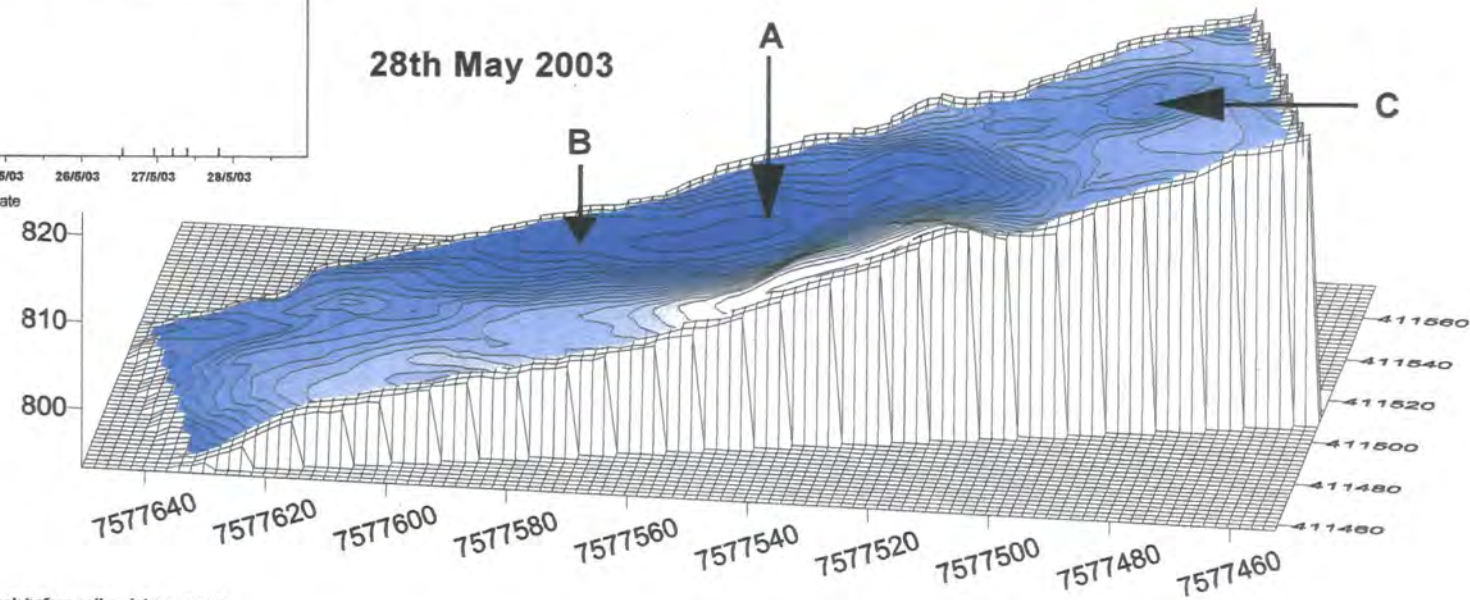
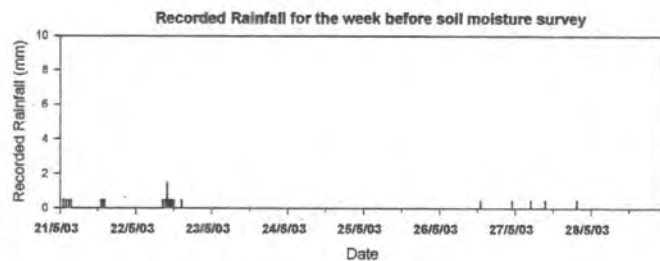
Figure 4.9 (a) shows the soil moisture distribution across the Hillslope Grid for 28<sup>th</sup> May and 2<sup>nd</sup> June 2003. The recorded rainfall for the week before the May survey shows that a number of small rainfall events occurred. The day before the survey, around 1.5 mm of rain was recorded. The soil moisture surface shows a large, very wet area in the centre of the grid (labelled A), concentrated at the base of the central topographic 'dip'. The wet area extended into the large flat area in the northern half of the grid (B). In the southern (upslope) area, a region of wet soil can be seen in a flat area before the first 'dip' (C). A smaller wet area can also be seen at the base of the Hillslope Grid.

Similarly, the June precipitation record shows around 2 mm precipitation the day before the survey. The soil moisture survey data shows that the central wet area persisted (D), becoming localised in the topographic hollow and the eastern part of the grid.

In July (Figure 4.9 (b)), the rainfall record shows that a number of rainfall events occurred leading up to the survey date, with a large event (4 mm) 3 days before and persistent rain on the day before. The grid is, however, drier than in the May and June surveys. Wet areas can be distinguished in the first (southern most) topographic hollow (A) and also at the base of the central 'dip' (B). There appear to be a number of linear features running in a north-south direction down the grid (C).

The central wet area extends to the north (downslope), in a line across the flat area to the east of the grid. On the western edge of the grid, a wet area was observed that extended downslope to the top of the northernmost 'dip' (D).

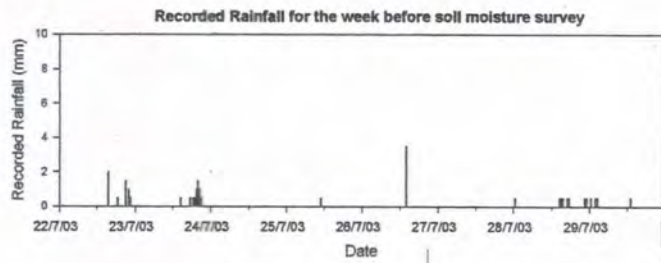
The September survey data does not show any of the north-south linear features observed in July. The rainfall record indicates only a few small events in the week before the survey. An area of elevated soil moisture was observed in the first (upslope) topographic hollow on the western edge of the grid (E). The other main feature observed was in the central flat region where an area of higher soil moisture was present (F).



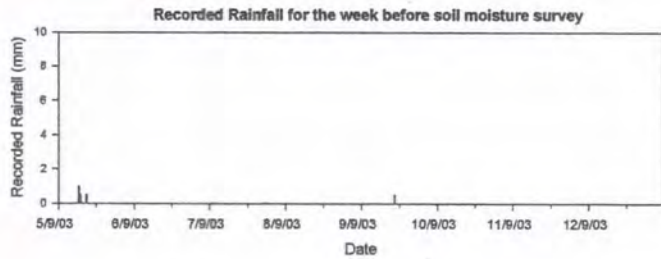
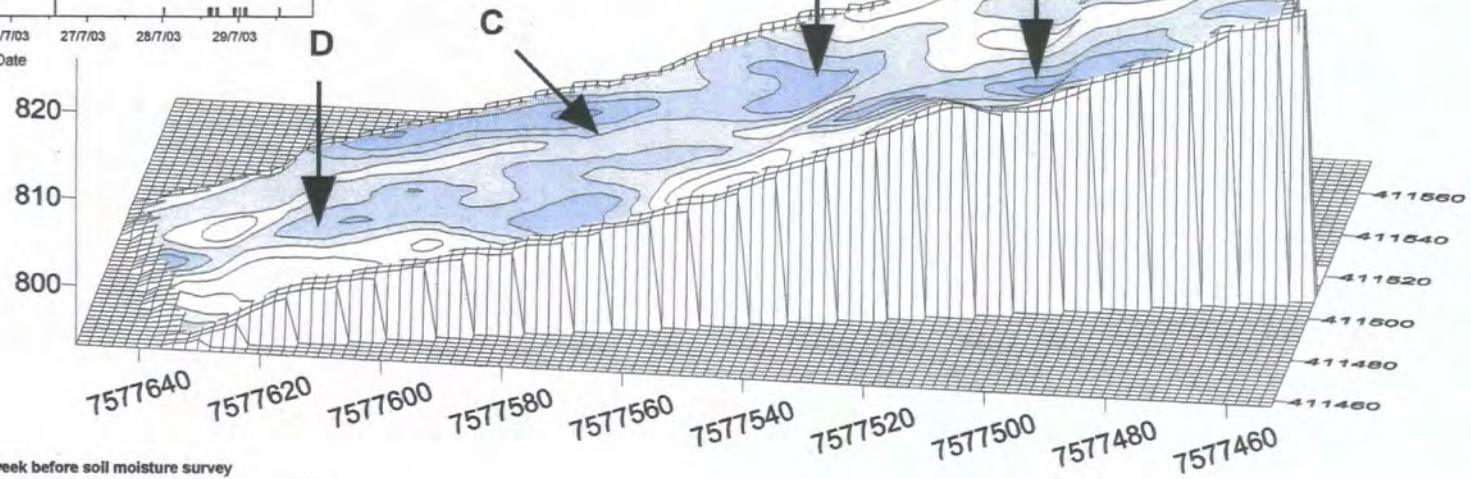
Soil Volumetric  
Water Content  
(fractional)

Figure 4.9 (a) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Hillslope Topography

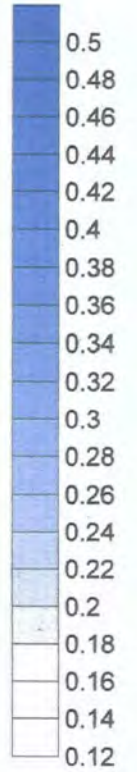
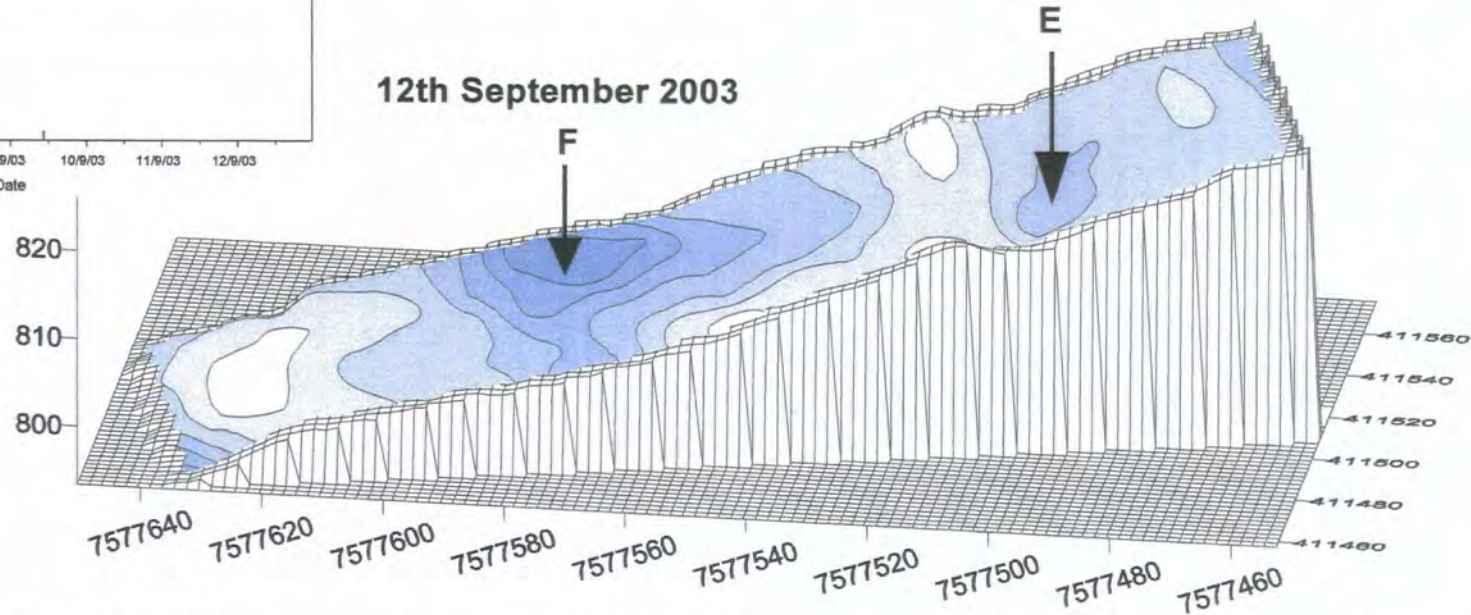




29th July 2003



12th September 2003

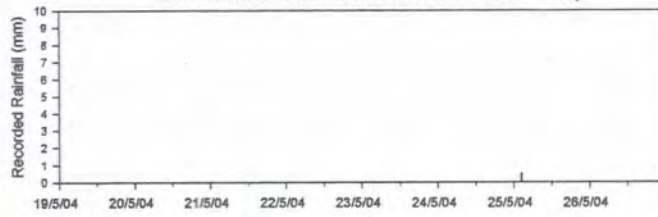


Soil Volumetric  
Water Content  
(fractional)

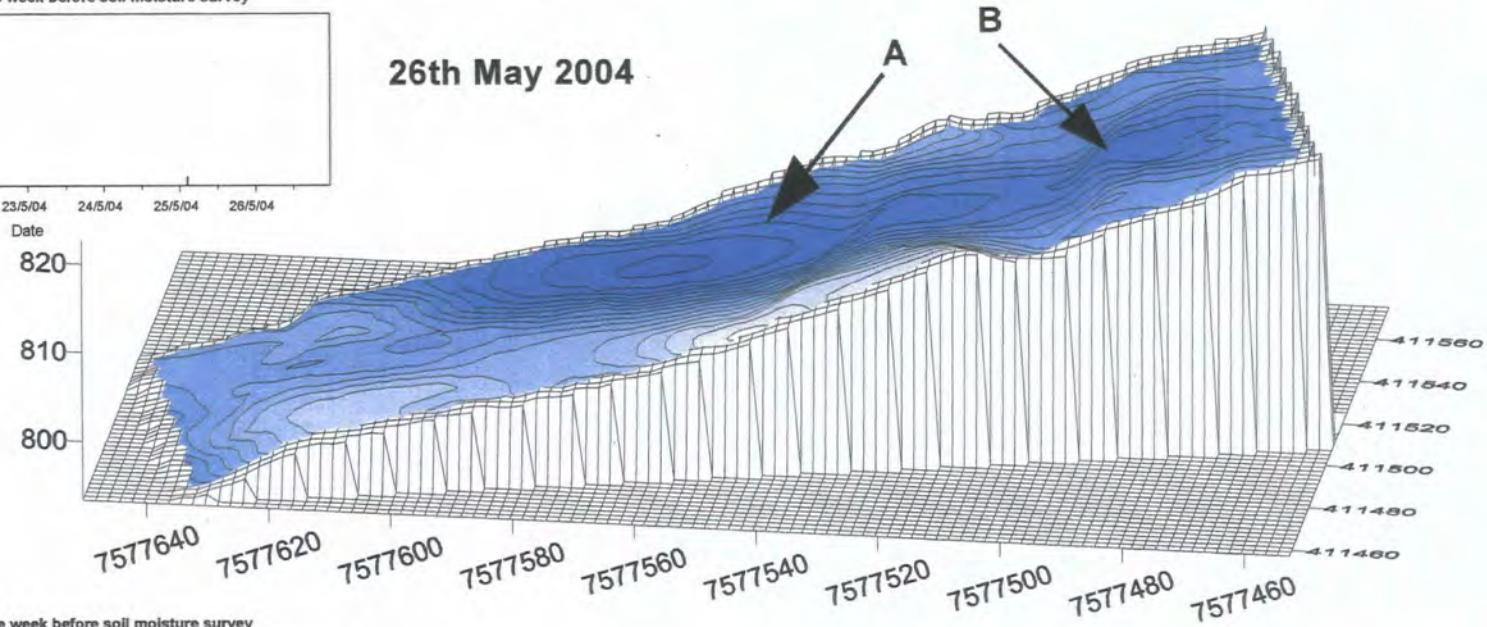
Figure 4.9 (b) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Hillslope Topography



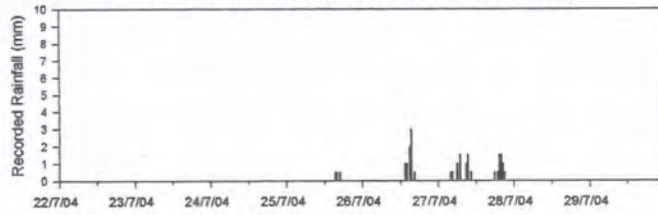
Recorded Rainfall for the week before soil moisture survey



26th May 2004



Recorded Rainfall for the week before soil moisture survey



29th July 2004

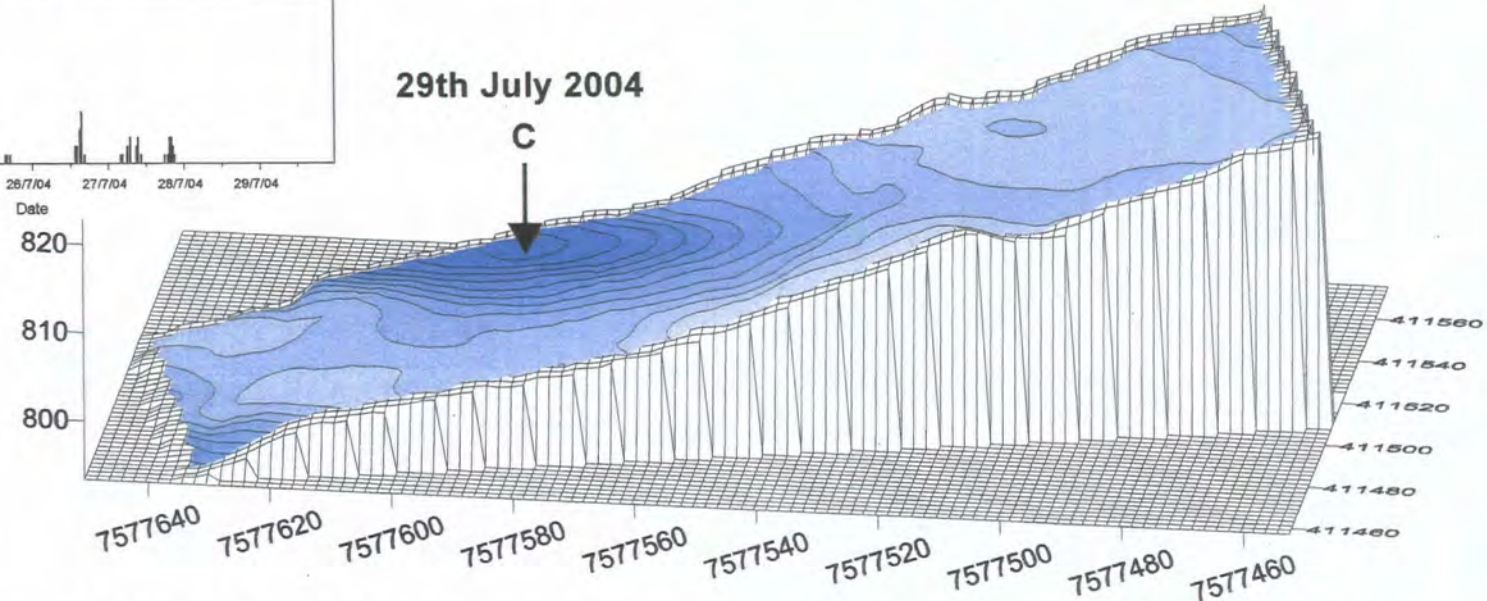
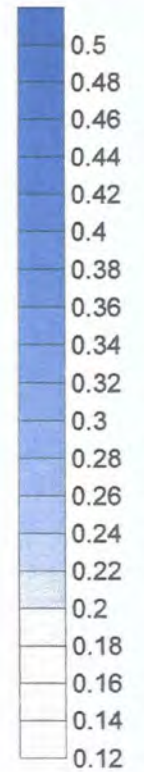
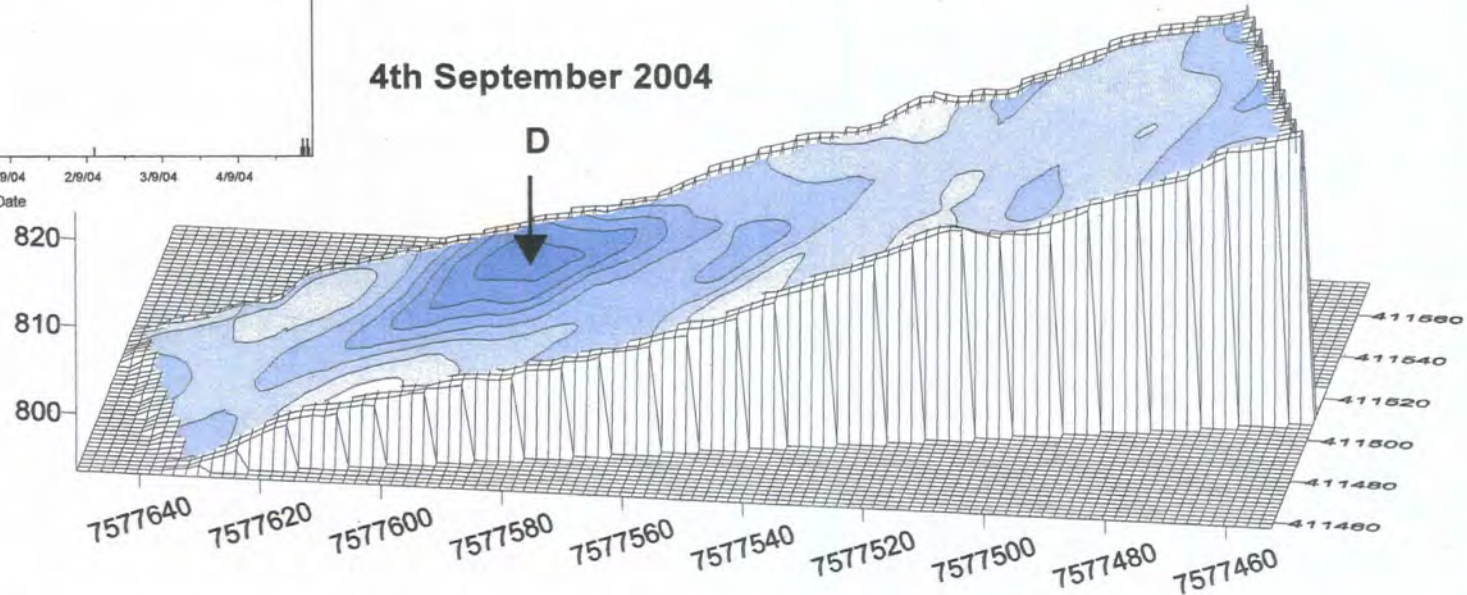
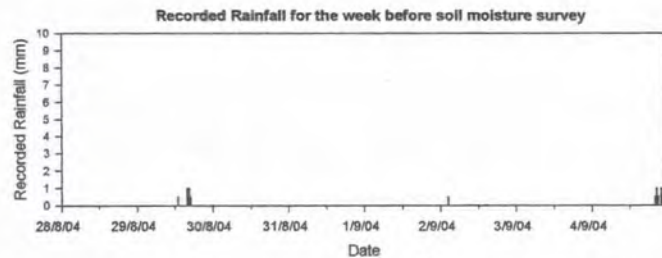
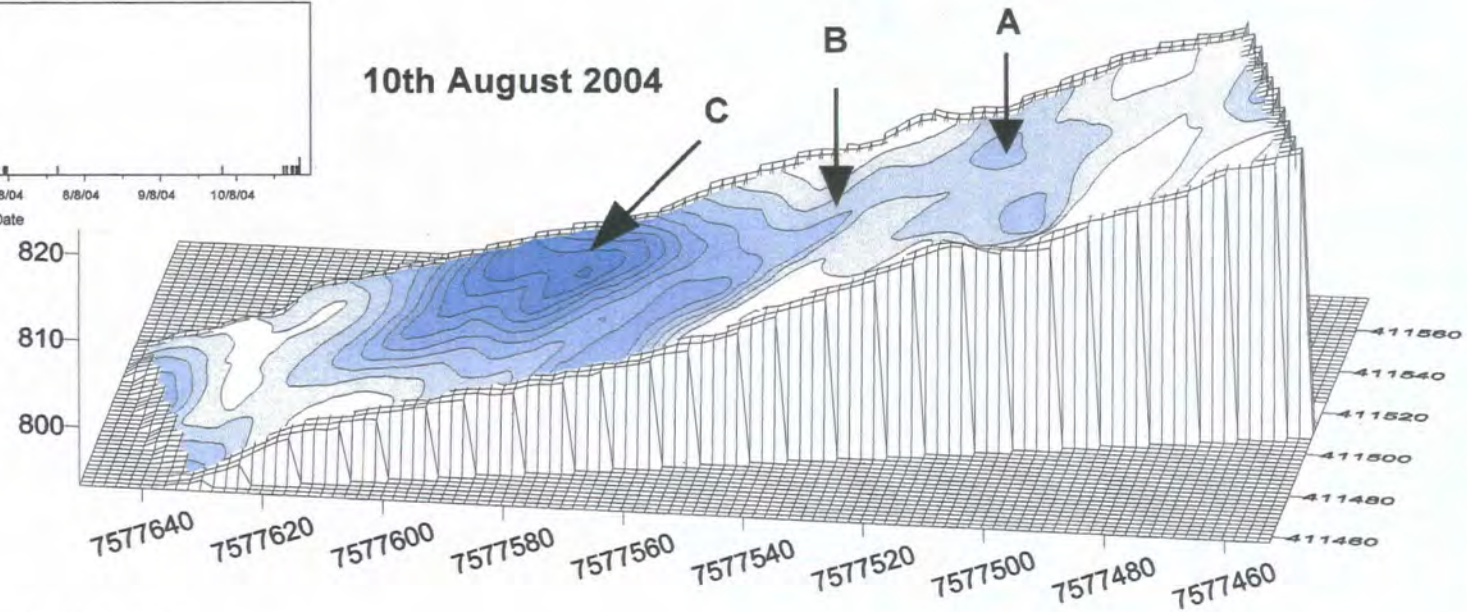
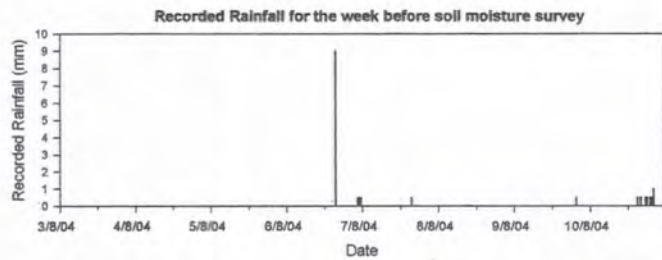


Figure 4.9 (c) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Hillslope Topography





Soil Volumetric  
Water Content  
(fractional)

Figure 4.9 (d) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Hillslope Topography



In May 2004 (Figure 4.9 (c)), the soil moisture distribution was very similar to the May 2003 survey. Only one small rainfall event was recorded but the grid is very wet. The large wet area again occurs in the centre of the grid (A) but this can be traced further upslope than in the previous year (B), passing through both the first topographic hollow and the central 'dip'. The July survey of 2004 is generally wetter than that of the previous year and the north-south structures are absent. The wet area in the flat region of the grid persisted (C).

Figure 4.9 (d) shows the soil moisture distribution across the grid for August and September 2004. The August survey shows an interesting distribution when observed with the topographical representation. At the base of the first (upslope) dip, two areas of elevated soil moisture can be distinguished (A). A linear wet feature appears to run from this region downslope into the next, central topographic hollow (B). This drainage feature then leads to the wet area in the flat region (C). The September survey data shows the wet area persisting in the flat region of the grid (D).

#### 4.2.1.2 The Windfetch Grid

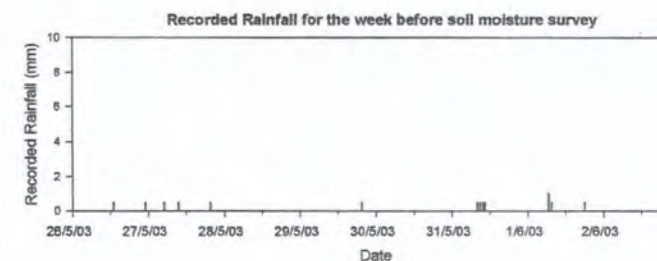
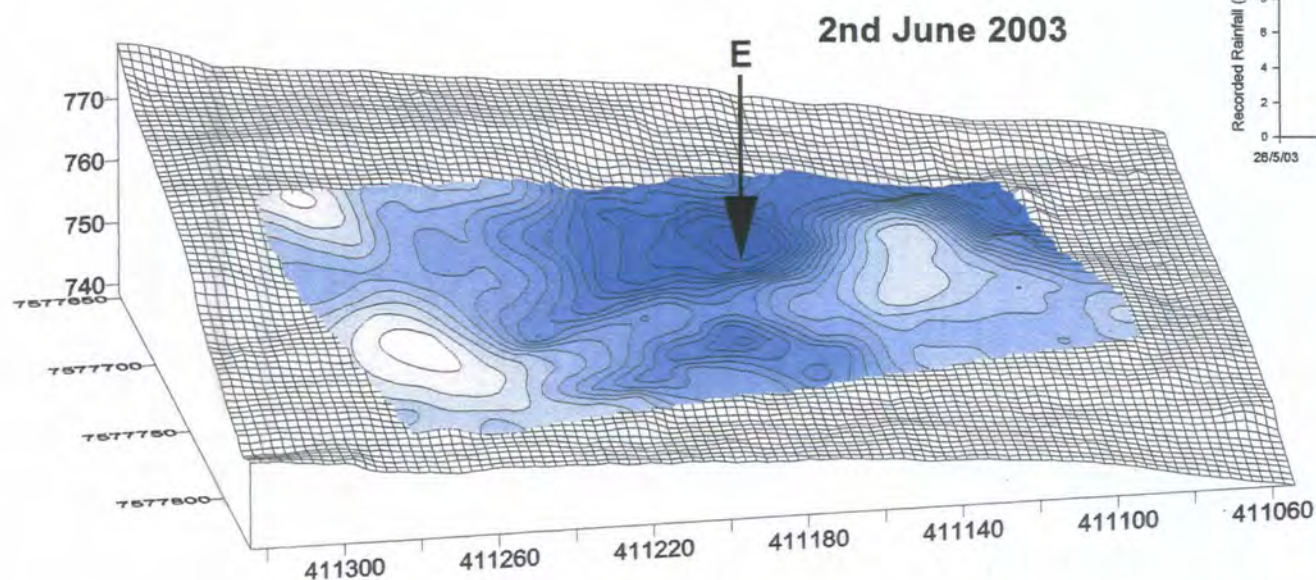
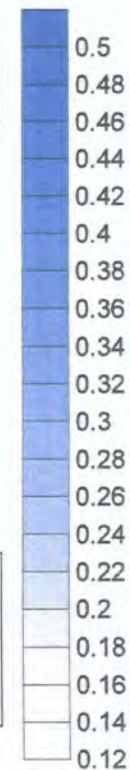
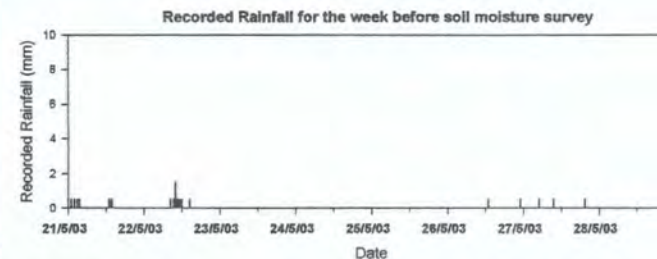
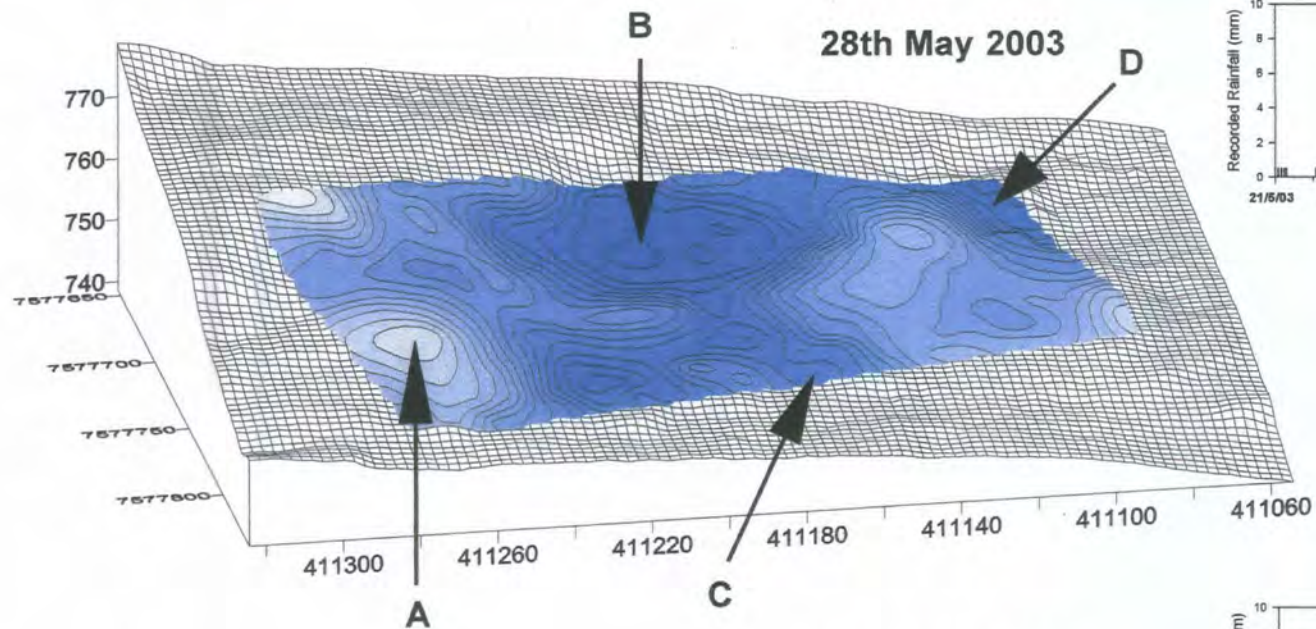
On the Windfetch Grid, Figure 4.10 (a) shows the soil moisture distribution for May and June 2003. In May, two areas of low soil moisture can be distinguished on the eastern edge of the grid (A). These coincide with two raised topographic features at these locations. The wettest areas occur in the central part of the grid, at the base of a slope (B). This wet area extends northwards with two wet areas identified on the northern edge of the grid (C). A second wet area was observed in the south-west corner of the grid (D). In June, the two drier areas on the eastern edge persist, as does the wet area at the base of the slope in the central region of the grid (E).

The July and August survey data are shown in Figure 4.10 (b). The Windfetch Grid appears much drier in these months; the dry areas on the eastern edge were present in both July and August but the wet area at the base of the slope is absent. There are few spatial soil moisture structures seen.

However, in September 2003 (Figure 4.10 (c)), an area of elevated soil moisture was observed in the central region of the grid (A). This extended eastward along the base of the slope. The two regions of dry soil were again present on the eastern edge of the grid.

The May and July 2004 surfaces (Figure 4.10(d)) show the area of wet soil occurring again at the base of the slope in the central region of the grid (extending to the north in the July survey) (A). The wet areas on the western edge of the grid were again present (B).

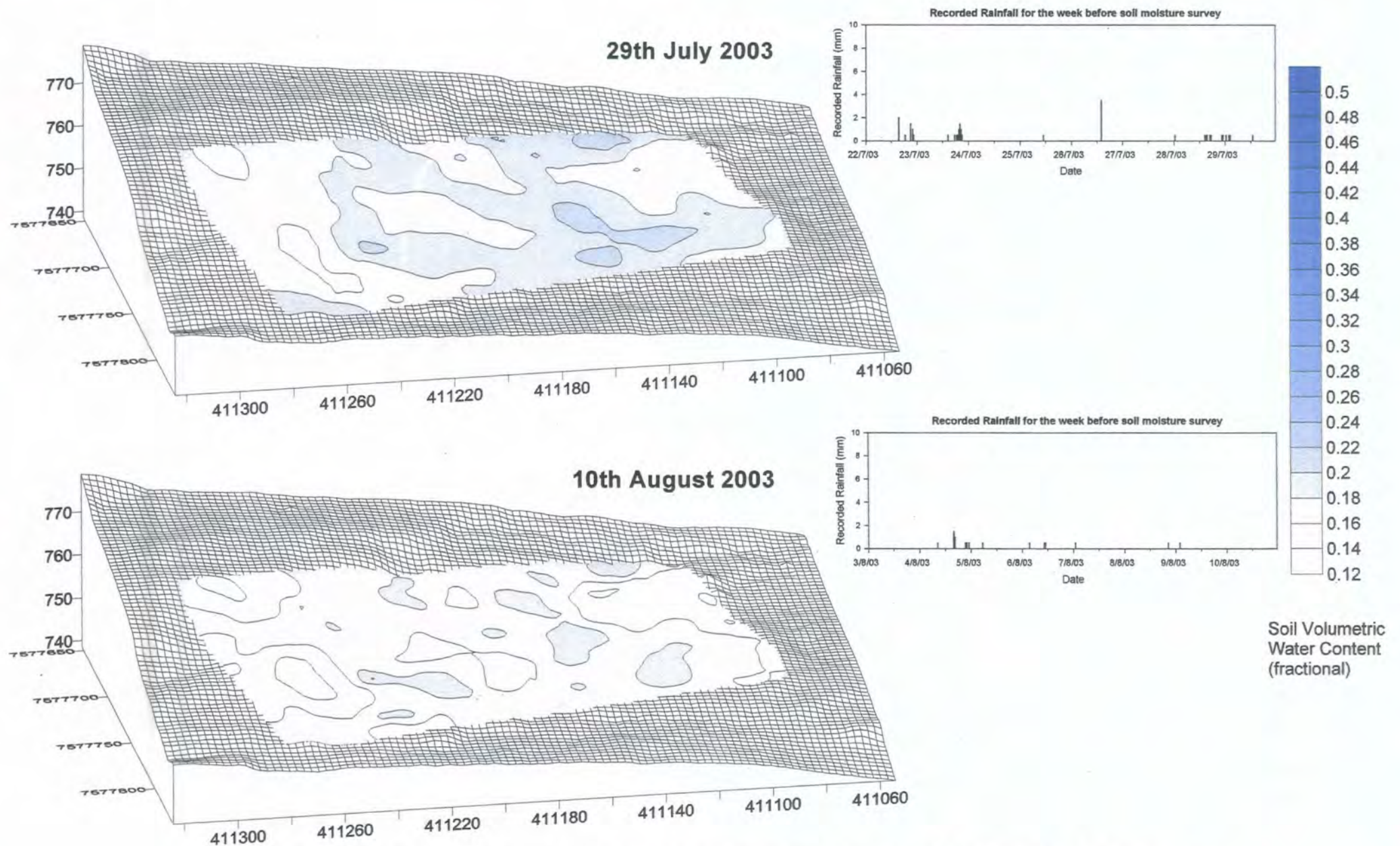
Figure 4.10 (e) shows the soil moisture maps for August and September 2004. These appear generally drier than the May and July soil conditions, however spatial soil moisture structures can be observed. In August, the wettest area of the grid occurred along the base of the slope in the central (southern) region (A). This pattern can also be seen on the September survey plot (B), along with many of the smaller spatial patterns of elevated soil moisture.



Soil Volumetric  
Water Content  
(fractional)

**Figure 4.10 (a) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Windfetch Topography**





**Figure 4.10 (b) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Windfetch Topography**



13th September 2003

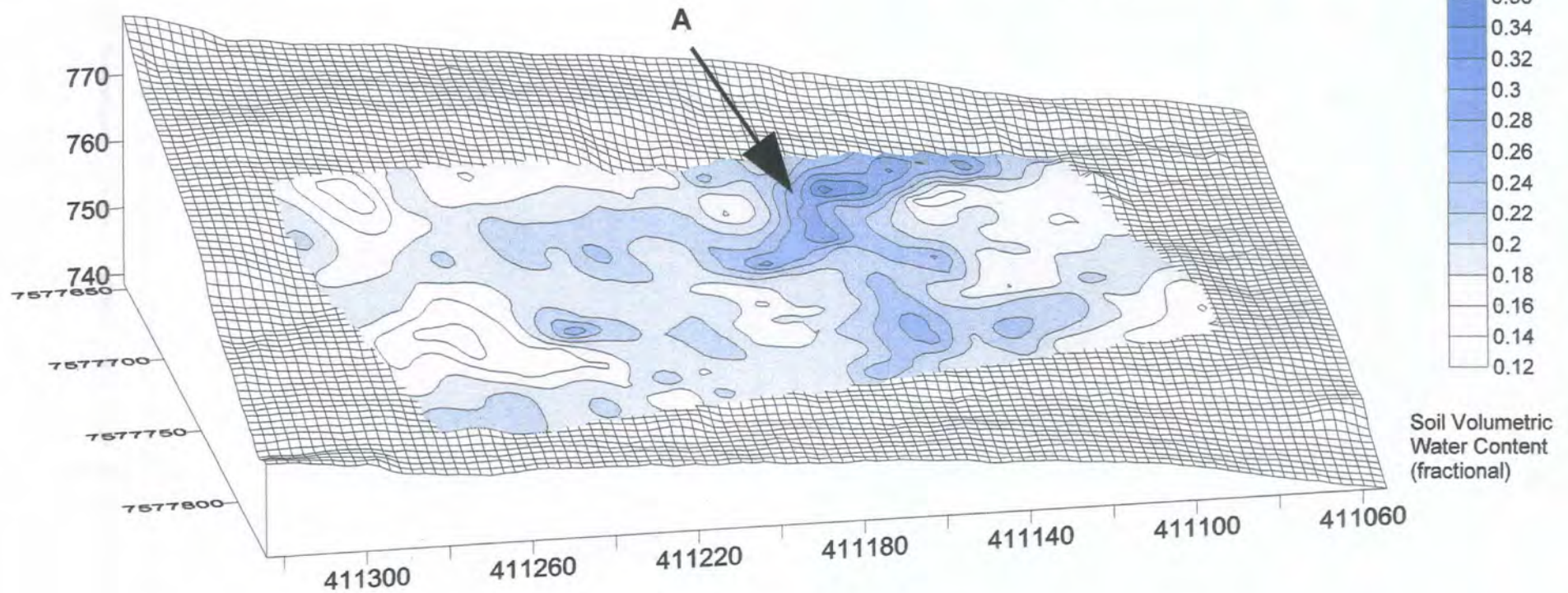
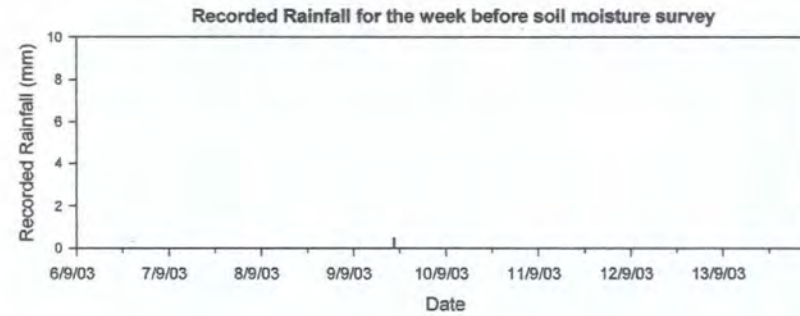
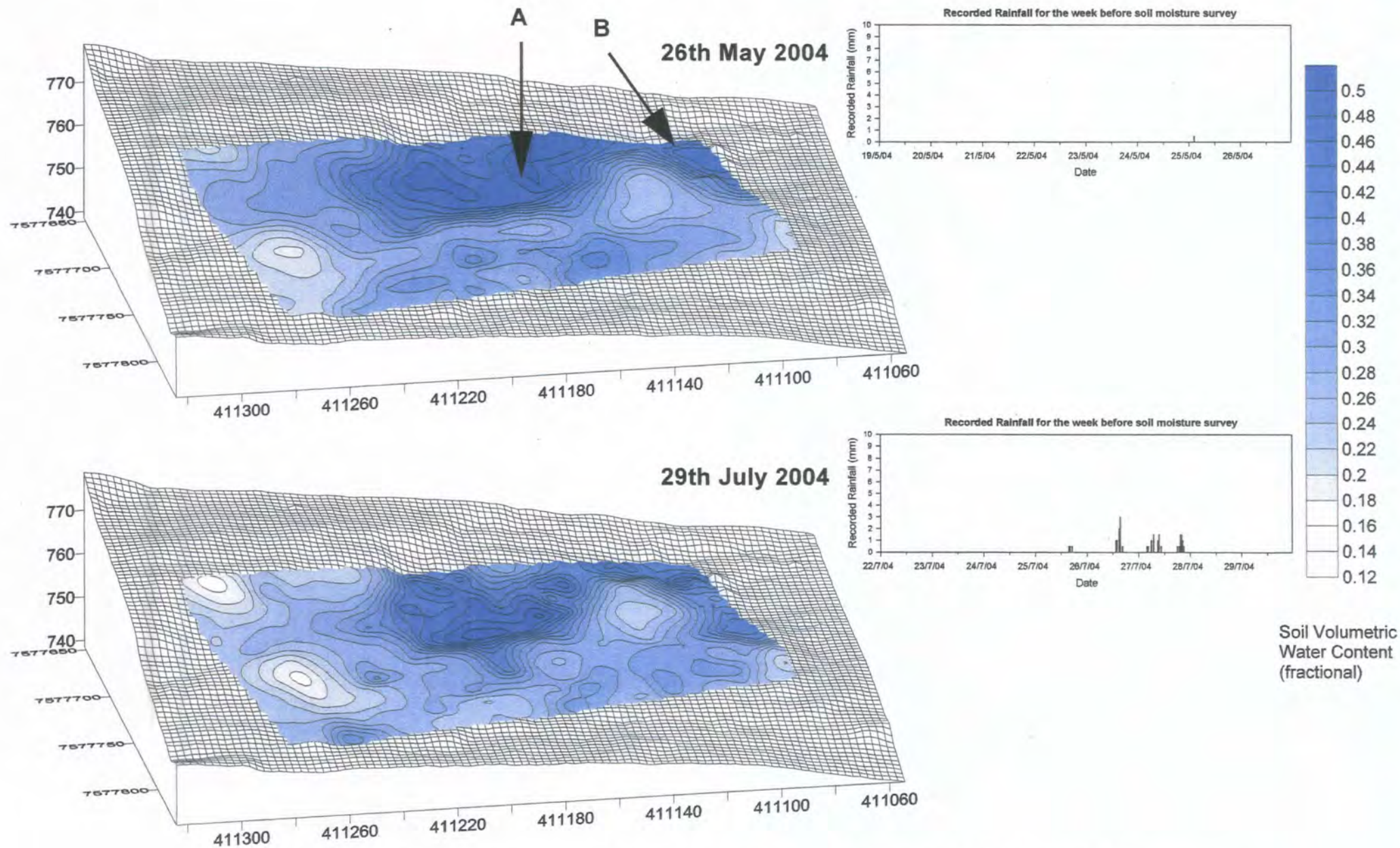


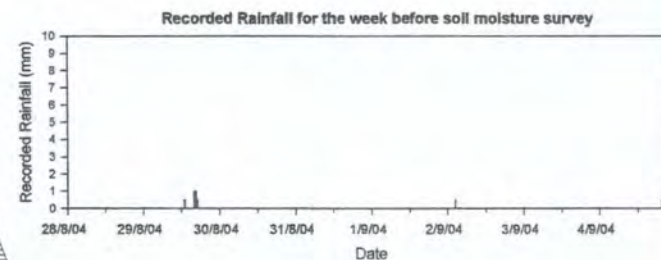
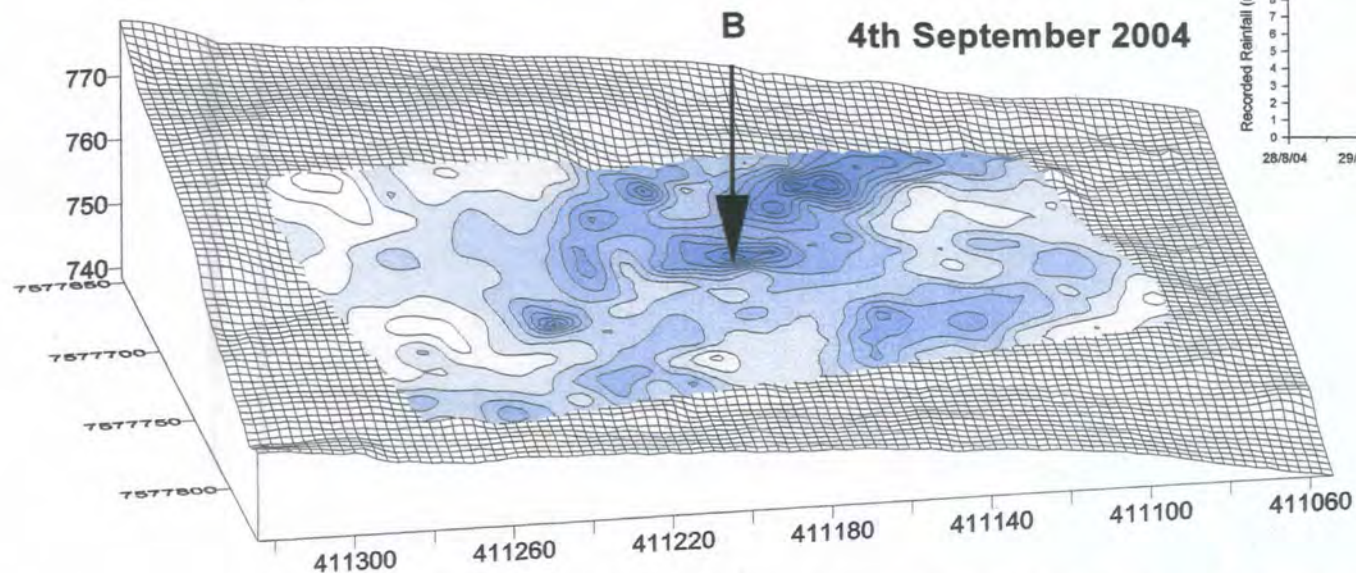
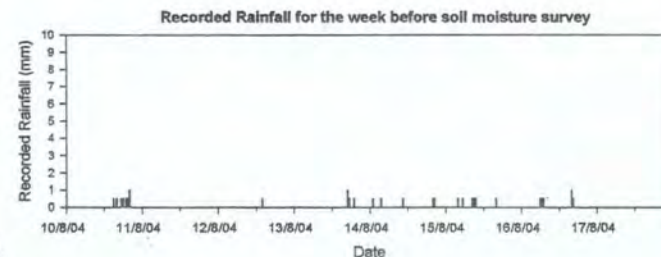
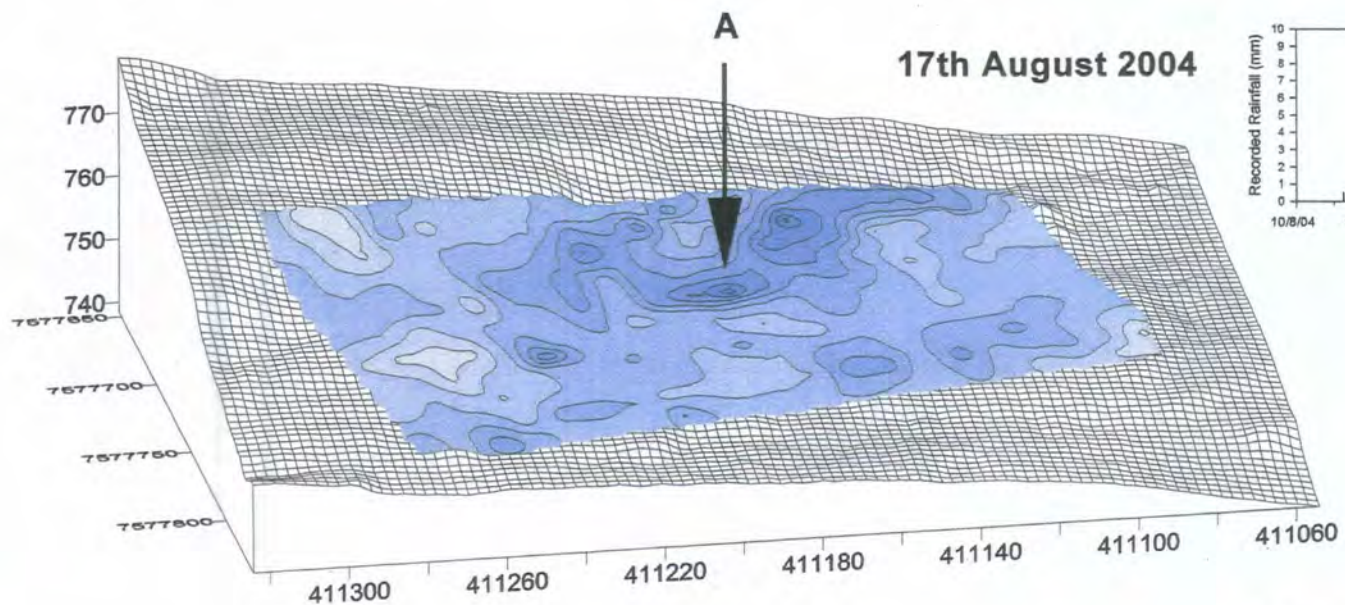
Figure 4.10 (c) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Windfetch Topography





**Figure 4.10 (d) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Windfetch Topography**





Soil Volumetric  
Water Content  
(fractional)

**Figure 4.10 (e) Soil Volumetric Water Content (5 cm depth) Surfaces Mapped onto the Windfetch Topography**

#### 4.2.2 Prediction of Wet Areas

In order to explore the prediction of wet areas using a terrain index, soil moisture was correlated with the upslope contributing area and the slope angle. The Topographic Index (Beven & Kirkby, 1979) approach uses the equation:

$$TI = \ln \frac{a}{\tan \beta}$$

where  $a$  is the area draining through a point from upslope and  $\tan \beta$  is the local slope angle (Beven, 1997). This is used as an index of hydrological similarity; areas with the same topographic index being assumed to be hydrologically similar. A location with a high topographic index value will saturate first.

The hillslope and Windfetch digital elevation models were used as inputs into a program called GRIDATB developed by Beven which had been modified by Dr. Douglas Clark (CEH Wallingford) (unpublished working code, Clark, pers. comm.). The output of this program was a grid of topographic index values that was then displayed using the ArcMap GIS Software (ESRI, 2004) which can be seen in Figure 4.11 (overleaf).

On the Windfetch Grid, Figure 4.11 shows an area of high topographic index running approximately east-west in the central (southern) region. This coincides with the base of the slope and also the soil moisture distributions shown in Figure 4.10.

The hillslope topographic index values suggest that the wet areas should occur at the base of the areas of high slope angle (the red areas shown on the output in Figure 4.11). There are also areas of high topographic index that occur on the large flat area in the northern half of the grid.



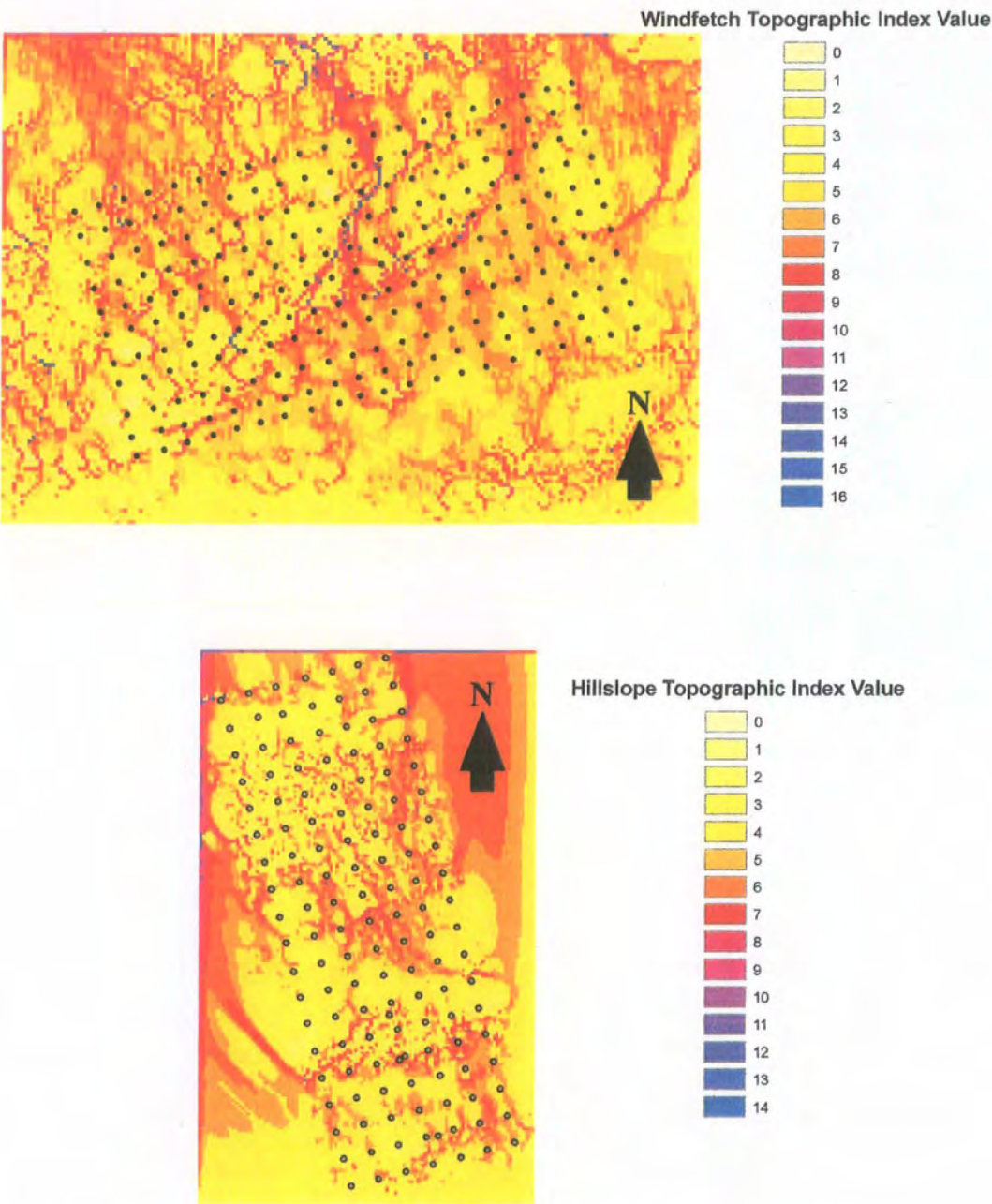
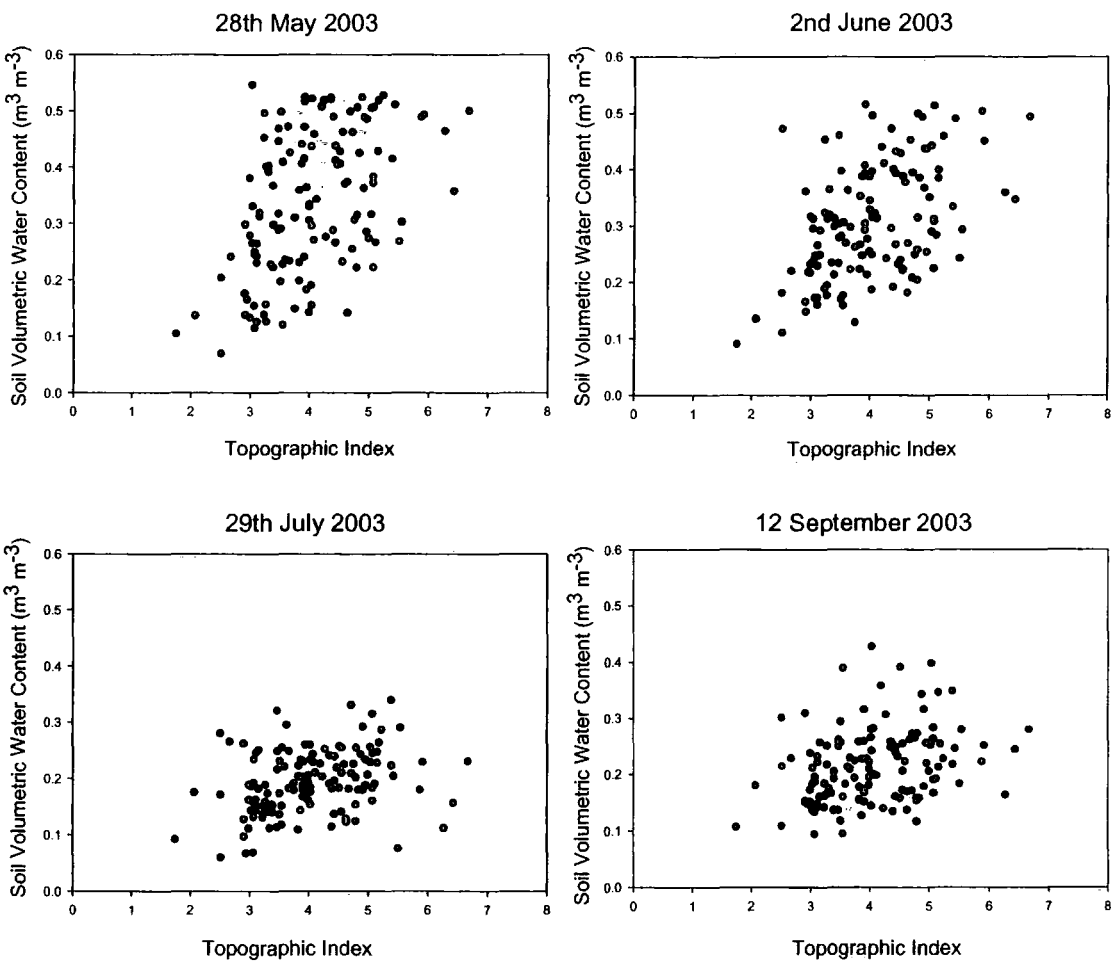


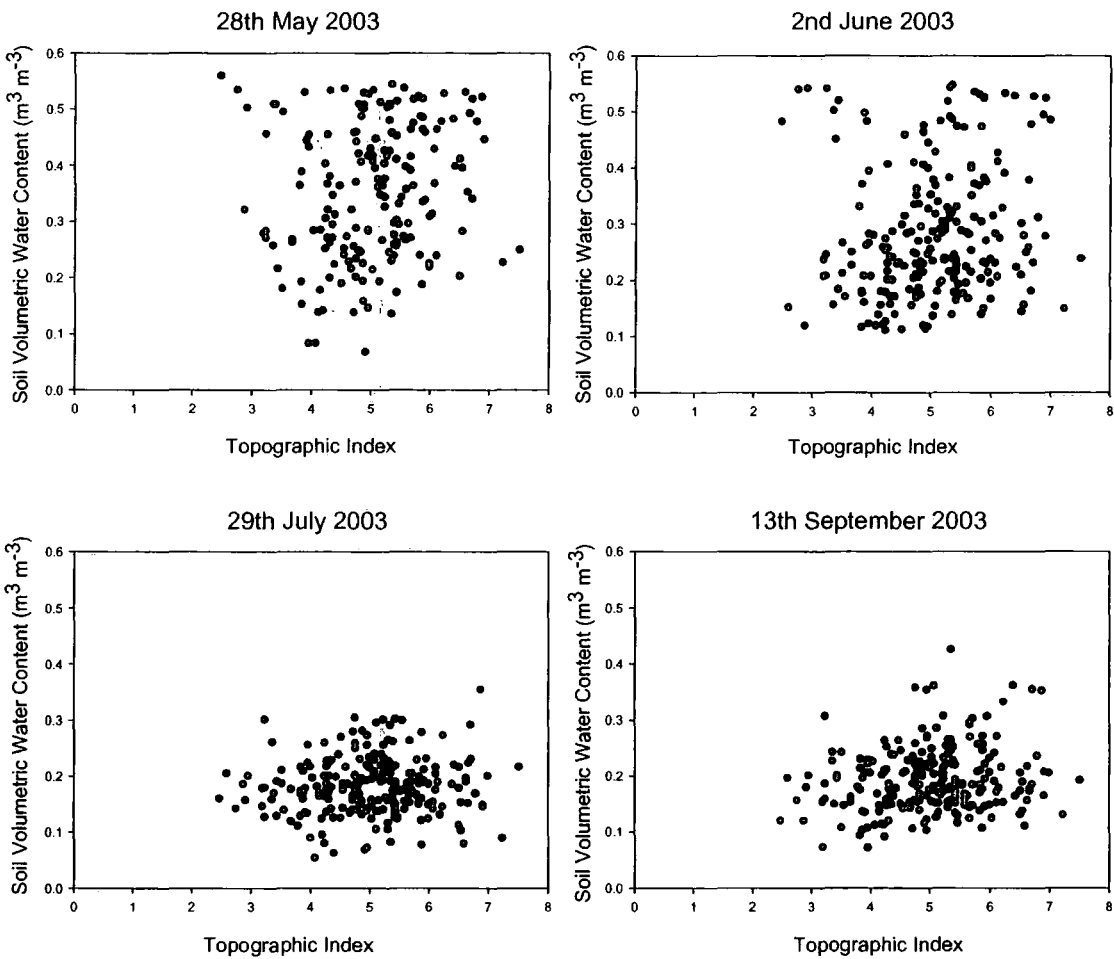
Figure 4.11 hillslope (1 m resolution) and Windfetch (1.5 m resolution) Topographic Index values

The observed soil moisture values obtained by the spatial surveys were then compared to the calculated topographic index at each grid point across both grids. The topographic index values were first smoothed (using the ‘blockmean’ command in ArcInfo GIS) to give average values for the grids at 5 m resolution. This smoothing was undertaken so that soil moisture values that were obtained at 10 m resolution could be compared with the topographic index values. At 5 m resolution, the topographic index values give a much more representative assessment of the processes seen at the study area. Graphs of observed soil moisture against topographic index were then produced for each of the spatial surveys. Figure 4.12 shows the relationship between observed soil moisture and topographic index for the Hillslope Grid for the 2003 surveys.



**Figure 4.12 Topographic Index (smoothed 5m values) and Soil Volumetric Water Content (SCIP surveys) for the Hillslope Grid during 2003**

Figure 4.12 shows there is only a slight positive correlation between observed soil moisture values and topographic index for the hillslope during the wetter months of May and June (just after the snowmelt). The drier summer months (July and September) are not correlated. The same analysis carried out for the Windfetch Grid (shown in Figure 4.13 below) shows that there is no correlation between the observed soil moisture values and topographic index over this flatter grid.



**Figure 4.13 Topographic Index (smoothed 5m values) and Soil Volumetric Water Content (SCIP surveys) for the Windfetch Grid during 2003**

The maps of topographic index displayed in Figure 4.11 indicate the areas that are prone to saturation first (those having high values) that appear to show some similarities with observed soil moisture distributions in the spring months. However, when compared to actual measurements, the topographic index values are not representative of the soil moisture state at the grid point measurement locations.

### 4.3 Chapter Summary

Using manual surveying techniques (Real Time Kinematic Surveying) and the acquisition of remote sensing (LiDAR) data, a high resolution digital representation of the study area was produced which allowed the physical features of the landscape (such as elevation, slope and aspect) to be described (Figure 4.5). Also, field observations and photographs have been used to provide evidence of micro-topographical features present on the hillslope which influence the distribution of water across the landscape during the snowmelt.

Slope maps (Figure 4.4) for the Hillslope and Windfetch sites derived from the topographic data highlight the ridges, hollows and terraces present at the study area. The average slope values for the Hillslope and Windfetch sites were approximately  $10^{\circ}$  and  $7^{\circ}$  respectively. The presence of localised areas of high slope are important when considering the soil moisture distributions since the topographic hollows present at the base of these slopes are sheltered from the effects of high winds and wind-blown snow. This means that larger shrubs (predominantly *Salix*) are present here and wind-blown snow has a tendency to accumulate and persist in these hollows. Also, the soil is deeper due to local deposition of leaf litter and the accumulation of wind blown soil.

On the hillslope, the wet areas appear at the base of the topographic hollows during the snowmelt. The soil water is then redistributed downslope, accumulating in an area of flatter topography. On the Windfetch Grid, the spatial patterns of high soil moisture occur at the base of the slope during the melt period. This suggests that the distribution of near surface soil moisture is topographically based during the spring period. The melt water from spatially distributed snow cover is redistributed downslope (due to persistent frozen ground below the snowpack) as lateral surface flow, accumulating in areas of flatter topography and thawed soil. However, as the season progresses, the Hillslope site shows that these areas of elevated soil moisture persist, while the Windfetch site shows considerably less spatial variation. This suggests that during wet periods, soil moisture variability is topographically driven, while during drier periods other factors (such as soil properties and vegetation structure) determine the soil moisture state.

The use of a terrain index to predict the soil moisture state due to topographic influence has been examined. Although maps of the Topographic Index (Figures 4.12 and 4.13) suggest that saturation should occur at the base of the slopes, direct comparison of observed soil moisture values and Topographic Index values showed that the two were not well correlated. However, correlation between topographic indices and observed soil moisture implies other independent variables remain constant across the study area. Topographic indices do not take into consideration factors that may be important in determining spatial soil moisture patterns, factors such as spatially heterogeneous soil types, soil depths and vegetation. Therefore, the lack of correlation between Topographic Index and observed soil moisture at the Abisko study sites may be due to a number of factors; the spatial variation in soil properties, soil depth and vegetation structure may all be contributing factors.

## **Chapter 5**

**To determine the relationship between snow depth, soil properties and vegetation height and their relation to soil moisture distributions**

The data presented in the previous two chapters has given an insight into the spatial and temporal soil moisture distribution and the links this distribution has with the topography of the site. Surveys were also carried out to quantify some physical aspects of the study area which would be relevant to the modelling process and help to build a picture of the hydrological processes operating at the fieldsite.

This chapter will present snow, soil and vegetation data measured at the fieldsite during the study period.

### 5.1 Field Methods and Data Collection

Throughout the study period comprehensive field notes and photographs were taken to document the observed hydrological processes occurring across the site. These included photographs taken in late winter and early spring that show the snow cover across the landscape at these times.

Comprehensive sampling of spatial snow patterns across the fieldsite was carried out in the scope of the STEPPS project but was beyond the scope of this thesis. However, to provide an insight into the spatial distribution of snow cover across the hillslope, two snow surveys were carried out on the 28<sup>th</sup> February and 3<sup>rd</sup> May 2003. Snow depth was measured along three transects by inserting a marked aluminium rod into the snowpack. The location of the survey points was recorded using a TS315 Total Station Theodolite (Trimble Inc.) which gave angle and distance measurements in a local 'grid'. The coordinates of the survey points were then derived using trigonometry to adjust the local grid to the WGS84 grid (GPS coordinates) (these calculations were undertaken with help from Andrew Wiltshire, University of Durham). The snow surveys were carried out before the establishment of the ten metre grid used for the spatial soil moisture surveys so there are slight deviations of the transects from the measurement points in this grid.

Soil profile data was provided from surveys carried out by Dr. Robert Baxter (University of Durham) and Dr. Philip Wookey (University of Uppsala, now at the University of Stirling)



on the 21<sup>st</sup> July 2003 (unpublished data, pers. comm.). Soil samples at four sites (*Empetrum*, *Salix*, *Sedge* and exposed ridge) were classified based on their texture, grain size, Munsell classification, pH, bulk density and loss on ignition. In addition to these surveys, the soil depth was measured over both the Hillslope and Windfetch Grids. A walking stick auger was used to measure the depth of the soil to the underlying rock, four depths being taken around each grid point (within a 50 cm radius of the point) and the measurements averaged to give a soil depth at that point.

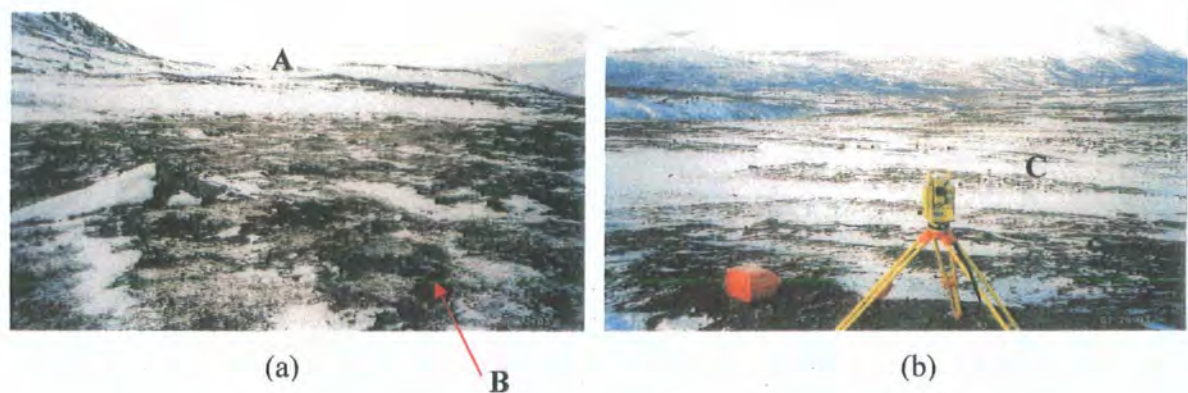
Surveys of the Hillslope and Windfetch Grids were also carried out to measure the vegetation height at each grid point. This was undertaken on the 3<sup>rd</sup> and 30<sup>th</sup> August 2004 and was achieved using a cane and measure to determine the height of the vegetation above the soil surface.

## 5.2 Analysis of Data

Data collected by manual surveys of both the Hillslope and Windfetch Grids are presented here. These data complement the soil moisture and topographic surveys carried out at the fieldsite and are relevant to this study, especially when considering the hydrological modelling that will be presented in Chapter 6.

### 5.2.1 Snow Observations and Surveys

Observations made at the fieldsite indicate that the snow cover is strongly influenced by the strong winds that frequently blow across the area. The snow accumulates in drifts in the topographic depressions and any snow present on the exposed ridges is quickly redistributed by the wind. Figure 5.1 shows the snow cover at the end of February 2003 on the Hillslope and Windfetch Grids. The Hillslope Grid clearly shows the snow accumulated in the topographic 'hollow' (A) and the flatter, more exposed area in the foreground (B) having only patchy snow cover. The Windfetch Grid photograph shows the difference in snow cover between the exposed ridges and topographic hollows, the ridges being relatively snow free (C).



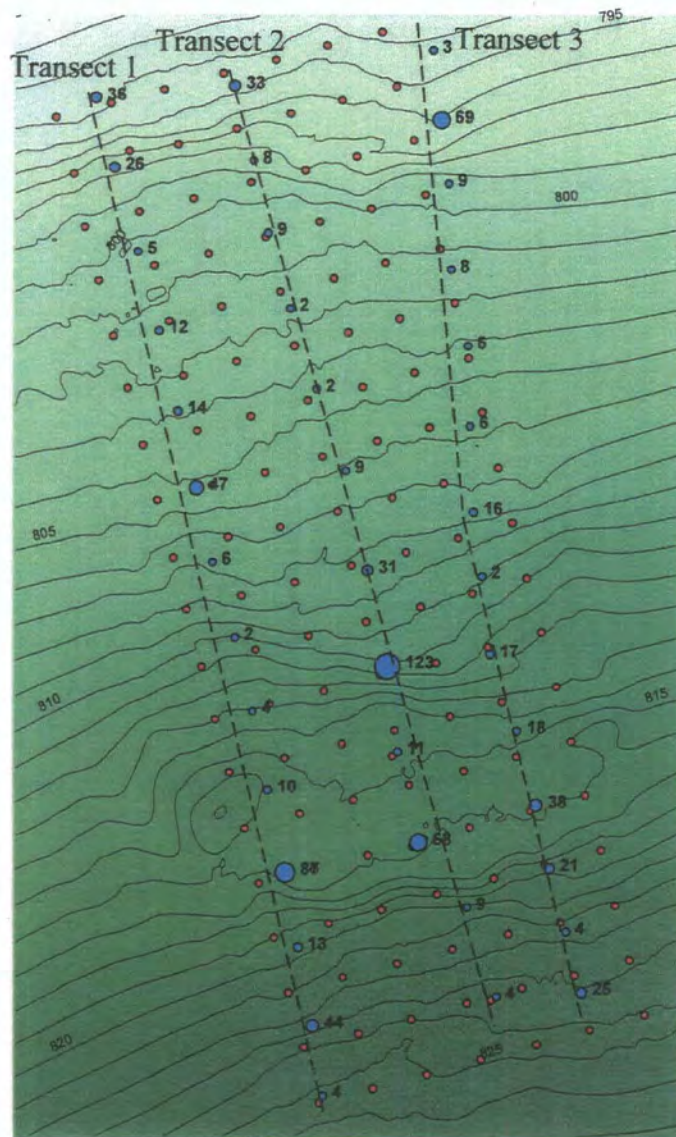
**Figure 5.1 Photographs of the Hillslope Grid (a) (taken 22<sup>nd</sup> February 2003) and the Windfetch Grid (b) (taken 26<sup>th</sup> February 2003) showing the snow cover at the fieldsite**

The snow surveys, undertaken across the Hillslope Grid in late February and early May 2003, provided a more detailed insight into the snow accumulation effect observed in the topographic ‘hollows’.

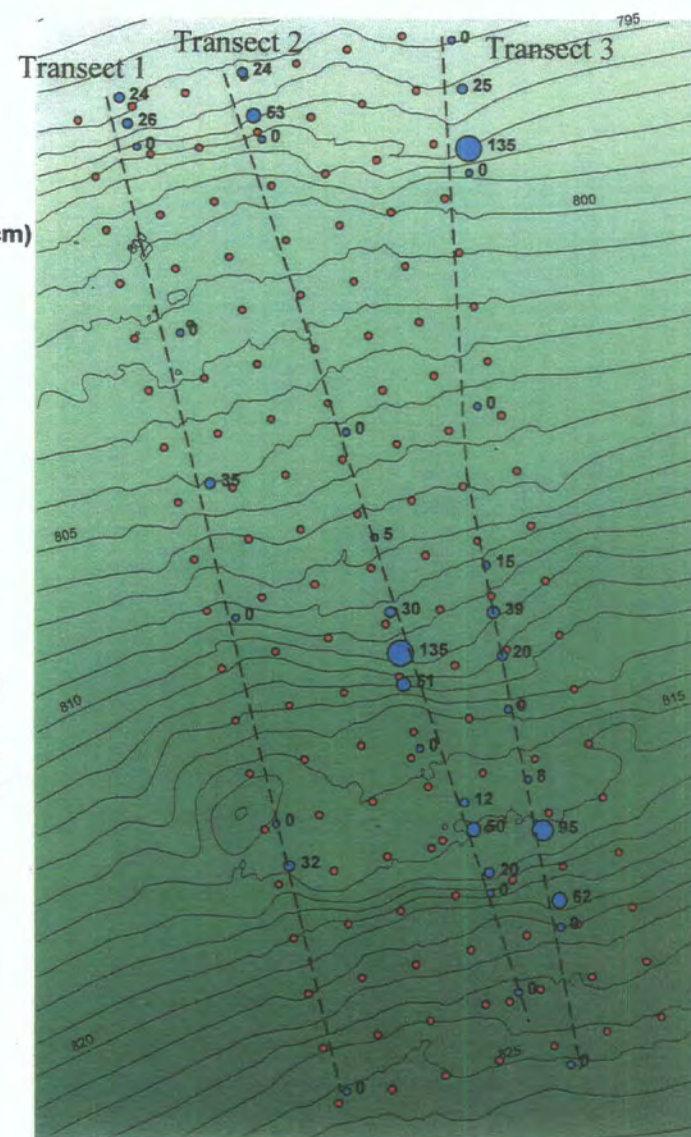
Figure 5.2 shows the location of transects and snow depth measurement points across the Hillslope Grid (displayed in ArcMap). The measurements made along the transects in February show that the snow depths across the Hillslope Grid range from 2 cm (in the exposed areas) to 123 cm (in the topographic ‘hollows’). The snow is seen to persist, with the same range of depths, into May.

To investigate the topographic controls of snow depth distribution for these two surveys, graphs of land surface elevation and snow depth were produced for each transect. These can be seen in Figure 5.3.





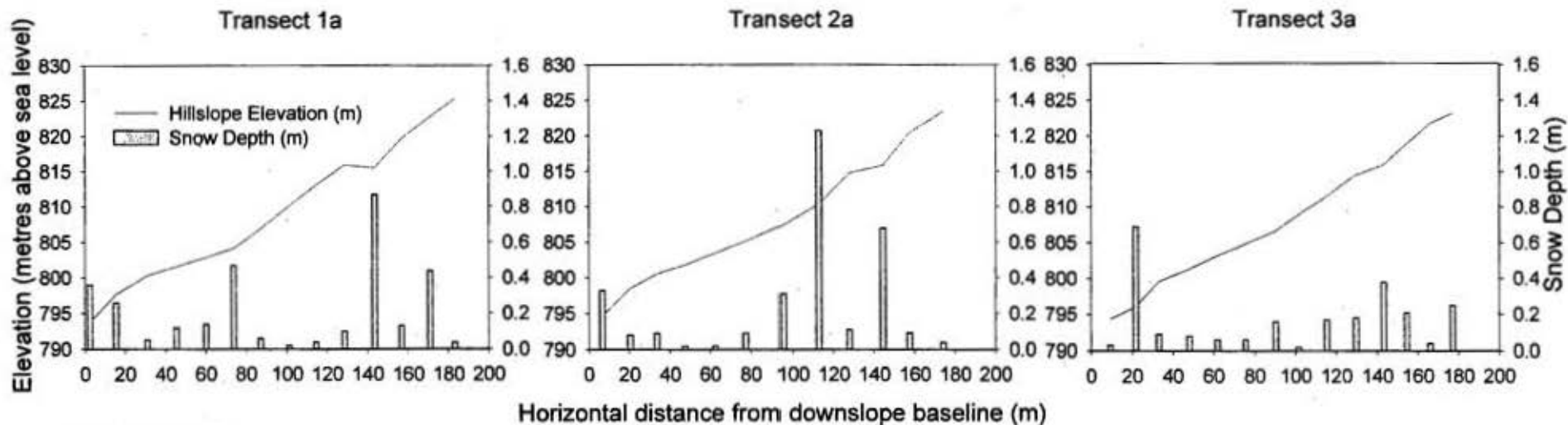
28<sup>th</sup> February 2003



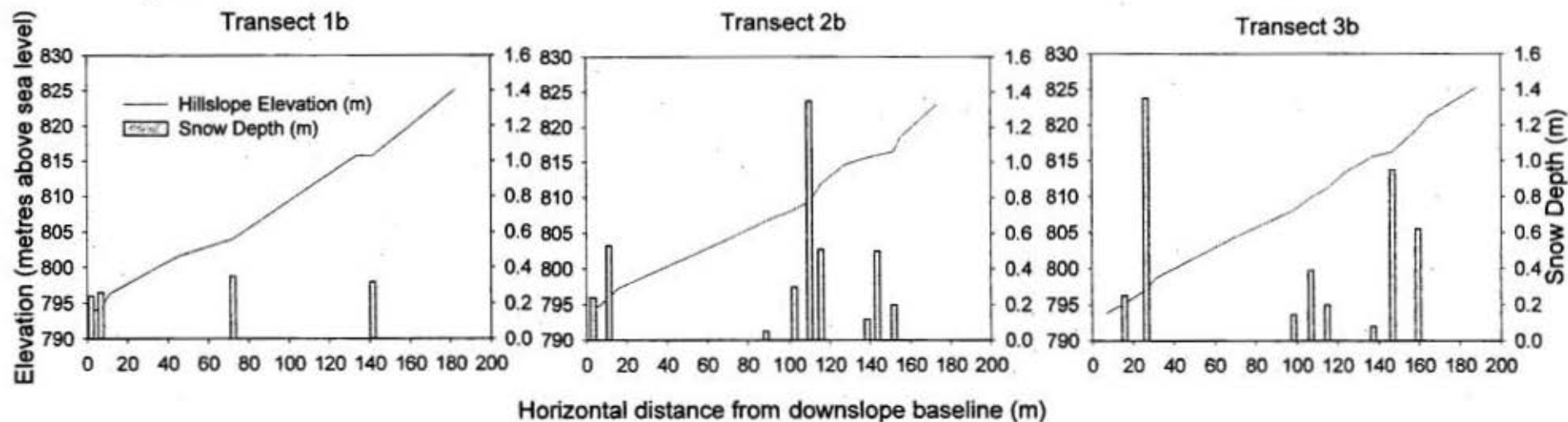
3<sup>rd</sup> May 2003

Figure 5.2 Snow survey data for the Hillslope Grid (displayed using ArcMap)

**28th February 2003**



**3rd May 2003**



**Figure 5.3 Snow depth and land surface elevation along transects of the Hillslope Grid for February and May 2003**

Although the transects were in approximately the same positions for the February and May surveys, the number of measurements taken and their location differ slightly between the two surveys. Figure 5.3 shows the hillslope elevation and snow depth against the upslope distance from a baseline at the northern edge of the Hillslope Grid (at the base of the slope).

In the February survey, the snow distribution along Transect 1a shows peaks at 2.5 m, 74 m, 144 m and 171 m. These coincide with the bases of the steeper slopes in topography shown by the hillslope elevation trace. A local anomaly is seen at 171 m where a peak in snow depth was observed without a corresponding 'dip' in topography. This may have been due to snow accumulation on the leeward side of a large rock or boulder at the time of the survey. Transect 2a shows the same pattern of elevated snow depth at the base of the steeper slopes, in this case at 6.6 m, 113 m and 144 m along the transect. The higher snow depths are again seen at the base of these slopes in Transect 3a at 22 m and 143 m. All of the transects in the February survey show that at all of the measurement points there was some snow cover, albeit not very deep in places.

In May, the persistent snow in the topographic 'hollows' is more distinct. Transect 1b shows snow cover at 2 m, 7 m, 72 m and 142 m. The topography of this transect shows the distinctive 'stepped' appearance of the hillslope, with the snow occurring at the bases of the steep slopes. This is repeated in transects 2b and 3b. Along transect 2b the snow depth peaks at 11 m, 110 m and 143 m. Transect 3b shows peaks in snow depth at 26 m, 107 m, 147 m and 159 m. The snow cover observed in the May survey only occurs in the topographic 'hollows'; no snow was measured along the flatter stretches of the hillslope.

The snow cover measured in the May survey was representative of the conditions observed during the spring snowmelt. The snow on the flatter, more exposed areas of the hillslope melted first, leaving the snow drifts that accumulated in the sheltered 'hollows' that persisted for up to a month after the initial thaw. This process was captured in a series of photographs, taken during the snowmelt period of 2003, of one such 'hollow' on the Hillslope Grid. Figure 5.4 shows the persistent snow on the day of the snow survey (3<sup>rd</sup>



May) and the subsequent melting of the snow, with vegetation beginning to show, later in May before the area becomes snow free in June.

This persistent snow cover late into the spring causes numerous problems when attempts are made at modelling the hydrological processes operating during the thaw period. These will be discussed in Chapter 6.



(a)



(b)



(c)

**Figure 5.4 Photographs of a topographic 'hollow' on the Hillslope Grid during the spring snowmelt period 2003, (a) 3<sup>rd</sup> May, (b) 15<sup>th</sup> May and (c) 7<sup>th</sup> June**

### 5.2.2 Soil Surveys

Soil surveys were undertaken to determine the soil type of four plots representative of the main land cover types (*Empetrum*, *Salix*, *Sedge* and exposed ridge). This soil profile data was provided by Dr. Robert Baxter (University of Durham) and Dr. Philip Wookey (University of Stirling).

The *Empetrum* site showed rooting throughout with no clear horizon differentiation. Horizons  $O_i$  and  $O_a$  were identified. Some pockets of silty material were present at 20 cm depth and deeper. The bulk density of the soil increased with depth, from  $0.1045 \text{ g cm}^{-3}$  in the top 5 cm rising to  $0.2754 \text{ g cm}^{-3}$  below 15 cm. The loss on ignition was high, decreasing with depth, from 94.84% in the top 5 cm to 80.28% at 15-18.5 cm. The soil was 30 cm deep with large cobbles at the base and was described as being highly organic, moderately to highly acidic and well-drained.

The *Salix* soil showed coarse roots throughout. Horizons  $O_i$  (0-3 cm),  $O_e$  (3-6 cm),  $O_a$  (6-11 cm), OA/AO (11-25/26 cm) and C (25-34 cm) were identified in this profile. The bulk density increased with depth from  $0.1564 \text{ g cm}^{-3}$  at 0-6 cm depth to  $0.4644 \text{ g cm}^{-3}$  at 11-19 cm depth. The loss on ignition was high in the top horizons (80.28%) decreasing to 35.38% at 11-19 cm. The OA/AO horizon had a loamy, crumb structure with organic staining throughout. This horizon had fine sand content which increased with depth and would break easily when wet. The soil was 38 cm deep to the C horizon with unsorted material at the base and was described as a highly organic soil profile, moderately acidic.

The *Sedge* soil again showed rooting throughout, with no clear horizon differentiation. Horizons  $O_i$  (0-14 cm) and  $O_e$  (14-19 cm) were identified. The bulk density increased with depth from  $0.1436 \text{ g cm}^{-3}$  (in the top 5 cm) to  $0.4103 \text{ g cm}^{-3}$  (below 10 cm). Loss on ignition was again high (84.38%) and decreased with depth to 75.10% at 10-13 cm. The soil was 19 cm deep and highly organic and mildly to moderately acidic.



The exposed ridge soil was a lithosol with no horizon differentiation. The soil contained unsorted material ranging from silt and sand to large rocks and boulders. Loss on ignition of this soil was low 2.77% and the soil was around 22 cm deep.

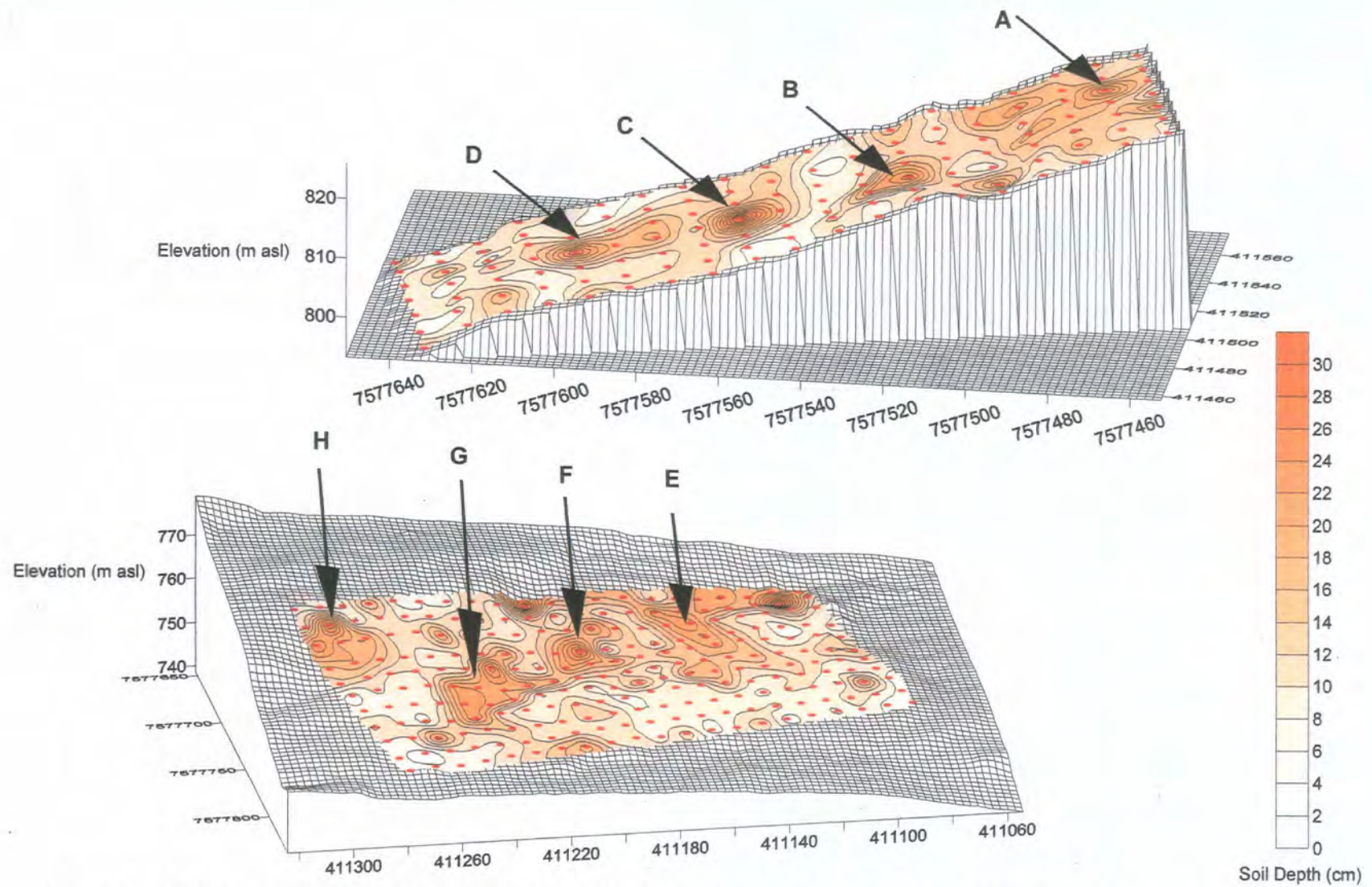
Table 2.2 (Section 2.4) gives a summary of the soil profile data including soil depth, pH, Munsell classification, bulk density and loss on ignition for the four land cover types.

In addition to the soil profile data, the soil depth was measured across both the Hillslope and Windfetch Grids (at each 10 m grid point). Figure 5.5 shows the averaged soil depth across both grids based on four readings at each point. The soil depth surfaces were created using the Surfer 7 (Golden Software, Inc, 1999) software package using a linear interpolation.

The soils of the study area were shallow, with an average depth across the Hillslope Grid of 10.9 cm and the Windfetch Grid of 10.4 cm. On the Hillslope Grid, the maximum soil depth recorded was 33.3 cm and the minimum was 4.0 cm. The maximum on the Windfetch Grid was 35.3 cm and the minimum was 4.3 cm.

On the Hillslope Grid (Figure 5.5), patches of deep soil can be seen in most of the topographic 'hollows'. At the top of the hillslope, deep soil was recorded (A) at the top of the first dip in topography and continued downslope into the 'hollow'. The second area of deep soil (B) occurs in the centre of the grid at the base of another dip in topography. The third area (C) again occurs in a 'hollow' and the deeper soil is seen to extend downslope to a fourth area (D) which occurs in the flatter region of the grid.

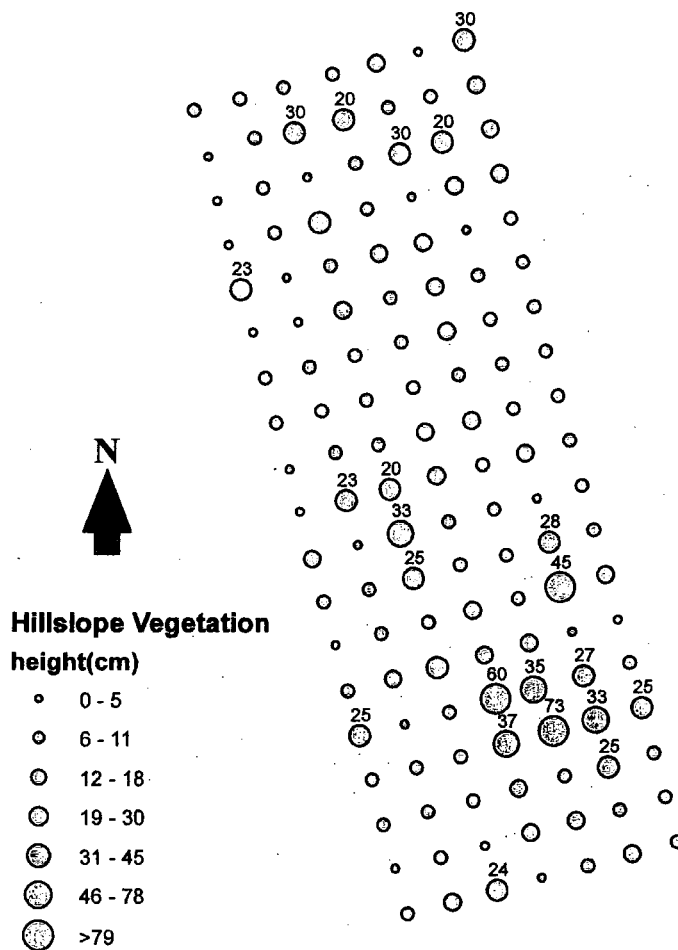
Four main areas of deep soils are observed on the Windfetch Grid, three of these occur along a line marking the base of the slope (E, F and G). The area of deep soil marked (H) in Figure 5.5 occurs in the south east corner of the grid in a small topographic 'dip'. The distribution of deep soils displayed on the Windfetch Grid show some similarities with the areas of elevated soil moisture shown in Figure 4.10 (e) for the 4<sup>th</sup> September 2004, occurring along the same line at the base of the topographic slope.



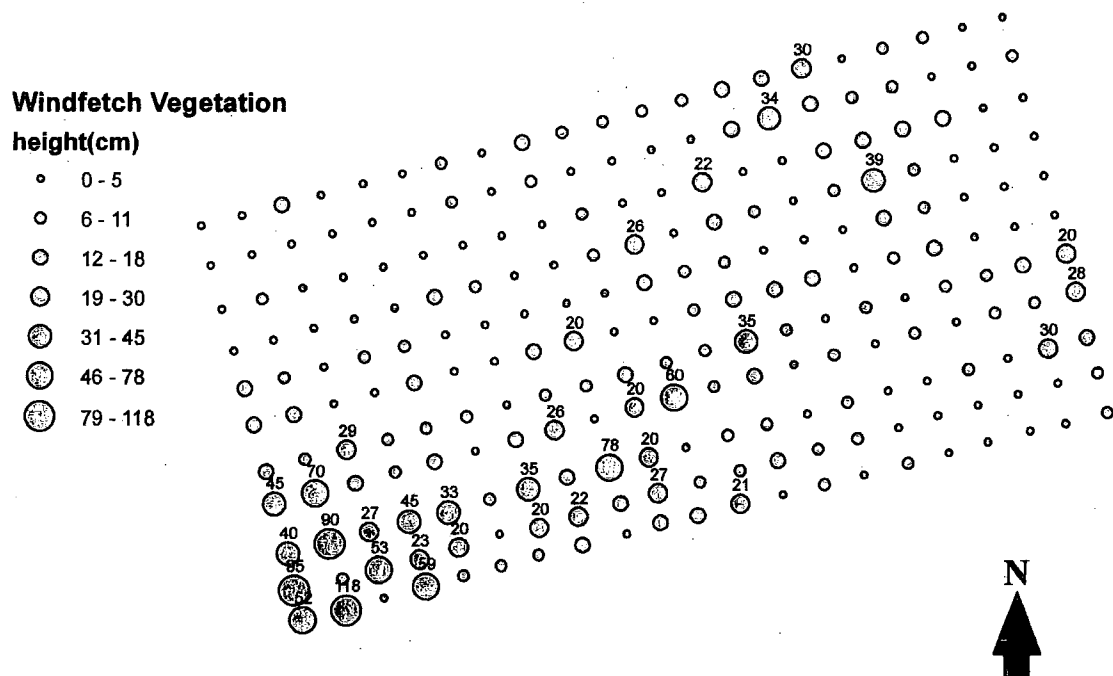
**Figure 5.5 Soil depths across the Hillslope and Windfetch Grids (based on an average of four measurements at each grid point)**

### 5.2.3 Vegetation Height

The vegetation height survey was carried out on the 3<sup>rd</sup> (Hillslope Grid) and 30<sup>th</sup> (Windfetch Grid) August 2004. Vegetation heights were measured at each grid point across both grids and the results are displayed in Figure 5.6 (a and b) for the Hillslope and Windfetch Grids respectively (displayed using ArcMap).



**Figure 5.6a Hillslope vegetation heights (3<sup>rd</sup> August 2004). Grid points are annotated where vegetation height exceeds 19 cm**



**Figure 5.6b Windfetch vegetation heights (30<sup>th</sup> August 2004). Grid points are annotated where vegetation height exceeds 19 cm**

Figure 5.6a shows the vegetation heights across the Hillslope Grid at the time of the survey. The tallest vegetation was 73 cm and occurred in one of the topographic ‘hollows’. All of the tallest vegetation (greater than 19 cm in height) occurred in the sheltered areas of the grid.

This pattern is displayed again on the Windfetch Grid (Figure 5.6b). Here the tallest vegetation was over a metre in height (118 cm) and occurred in the south west corner of the grid. A patch of tall vegetation was observed at this location. The pattern of tall vegetation appears to follow the base of the slope in this grid, in a south-west/north-east direction across the centre of the grid. These are also the areas observed to have the deepest soils.

### 5.3 Chapter Summary

Snow depth surveys were carried out in February and May 2003. These surveys confirm that the largest snow depths occur in the topographic hollows and that, more importantly, they persist in these areas into the spring thaw.

The soil moisture distributions described in Chapter 3 are clearly driven by the input of melt water during the spring thaw and it has been concluded that water from melting snow in the hollows is redistributed due to topographical lateral flow.

The accumulation of snow in the topographic hollows across the site is also linked to the patterns of vegetation in these areas. The tallest vegetation, measured in August 2004, occurs in the sheltered 'hollows' across both the Hillslope and Windfetch Grids. This effect is most prominent on the Windfetch Grid where the tallest vegetation occurs in the south-west corner of the grid and extends in a line across the grid in a north-easterly direction matching the base of the slope seen on this grid. The hollows typically contain the taller shrubs (*Salix*) since they are more sheltered and the more exposed areas of flatter topography are populated by smaller species such as *Empetrum* heath. The *Salix* in the hollows traps wind blown snow causing the deep snowdrifts. These snowdrifts insulate the soil; keeping the soil temperature at around 0 °C through the winter months.

Due to the heterogeneous nature of the landscape, the soil properties are considered to be highly spatially variable. The soils across the study area are highly organic and very shallow. The *Empetrum* soils are well-drained with a very high loss on ignition. The *Salix* soils are the deepest, displaying identifiable horizons down to the C horizon. The loss on ignition of these soils reduces with depth. The deepest soils tend to occur at the base of the topographic slopes in the sheltered areas of the site. Spatial measurement of soil depths across the Hillslope and Windfetch sites show that the deepest soils tend to occur at the base of the slopes and in the topographic hollows. This may be due to increased litter production from the *Salix* shrubs and also the accumulation of wind blown soil and soil creep. The soil depth will have a profound effect on the soil moisture status across the site,

the deeper soils providing more soil water storage but also saturating easily and providing lateral soil water movement compared to the well draining thinner soils.

Hence, the soil moisture distributions are greatly influenced by the heterogeneous snow cover, local soil properties and vegetation structure seen across the study area. During the thaw period, soil moisture variations are caused by the lateral redistribution of melt water from the snow in the hollows. In the summer months, soil moisture variability is predominantly determined by the spatially variable soil properties, soil depths and evapotranspiration from heterogeneous vegetation.

## **Chapter 6**

**To test the performance of a land surface model for modelling soil moisture in a  
heterogeneous Tundra landscape**



Meteorological and climate models are widely used to simulate global surface temperatures under different environmental conditions (such as increased carbon dioxide). Land surface schemes are used within these climate models to represent the land-atmosphere interactions at the land surface. The modelling of soil moisture and temperature are important elements of these models, especially for high latitudes where organic, seasonally (or permanently) frozen soils could become a source of carbon dioxide and methane should the temperature regime change.

Previous studies have highlighted shortcomings in the soil parameterisation of the Meteorological Office Surface Exchange Scheme (MOSES) for high latitudes (Harding *et al.*, 2002; Hall *et al.*, 2003; Hall, unpublished; Lloyd *et al.*, 1999).

In this chapter, the sensitivity of a land surface scheme to changing soil parameter values is explored using a Monte-Carlo parameter sweep method. This was undertaken in order to investigate the inclusion of organic soil parameter values which are not currently available in the chosen land surface scheme. The performance of the scheme during the soil freezing and thawing periods is then discussed with particular focus on the effects of snow cover, soil temperature and soil water phase.

A framework describing the modelling objectives and model runs is presented in Section 6.2, but first it is useful to provide a description of the land surface scheme used in this study.

### 6.1 Model Description

The MOSES (Meteorological Office Surface Exchange Scheme) land surface scheme was developed for the existing UK Meteorological Office (UKMO) climate model. MOSES provides a one-dimensional description of the exchanges of water, energy and carbon dioxide at the land surface and includes a ‘skin’ surface temperature, interactive surface resistance and soil water phase changes.

A comprehensive description of the MOSES model can be found in Cox *et al.* (1999) and the Hadley Centre Technical Note 30 (Essery *et al.*, 2001). Hall *et al.* (2003) provide a summary of the soil physics used to calculate soil water and temperature in the scheme which is relevant to this study and the processes are outlined below:

For the  $n$ th soil layer of thickness  $\Delta z_n$  (m) at temperature  $T_n$  (K), the heat balance is given by:

$$C_A \Delta z_n \frac{dT_n}{dt} = G_{n-1} - G_n - J_n \Delta z_n$$

(Equation 6.1)

where  $C_A$  ( $\text{J m}^{-3} \text{ K}^{-1}$ ) is the apparent volumetric heat capacity of the layer given by:

$$C_A = C_s + \rho_w c_w \theta_u + \rho_i c_i \theta_f + \rho_w [(c_w - c_i)T + L_f] \frac{\delta \theta_u}{\delta T}$$

(Equation 6.2)

Here,  $C_s$  ( $\text{J m}^{-3} \text{ K}^{-1}$ ) is the dry soil volumetric heat capacity,  $\rho_w$  and  $\rho_i$  ( $\text{kg m}^{-3}$ ) are the densities of water and ice,  $c_w$  and  $c_i$  ( $\text{J kg}^{-1} \text{ K}^{-1}$ ) are the specific heat capacities of water and ice,  $\theta_u$  and  $\theta_f$  are the volumetric concentrations of unfrozen and frozen soil water, and  $L_f$  ( $\text{J kg}^{-1}$ ) is the latent heat of fusion of water.

The diffusive heat fluxes into,  $G_{n-1}$  ( $\text{W m}^{-2}$ ), and out of,  $G_n$ , the  $n$ th layer are given by:

$$G = \lambda \frac{\delta T}{\delta z}$$

(Equation 6.3)

where  $\lambda$  ( $\text{W m}^{-1} \text{K}^{-1}$ ) is the local soil thermal conductivity.

The net heat flux advected from the layer by the moisture flux,  $J_n$  ( $\text{W m}^{-3}$ ) is determined by:

$$J = c_w W \frac{\delta T}{\delta z}$$

(Equation 6.4)

The soil hydrology component of MOSES is based on a finite difference approximation to the Richards' equation (Richards, 1931 cited in Essery *et al*, 2001). The prognostic variables are the total soil moisture content within each layer, given by:

$$M = \rho_w \Delta z \theta_s \{S_u + S_f\}$$

(Equation 6.5)

where  $\Delta z$  is the thickness of the layer and  $S_u$  and  $S_f$  are the mass of unfrozen and frozen water within the layer as a fraction of that liquid water at saturation. These are determined by:

$$S_u = \frac{\theta_u}{\theta_{sat}}$$

(Equation 6.6)

and

$$S_f = \frac{\rho_i}{\rho_w} \frac{\theta_f}{\theta_{sat}}$$

(Equation 6.7)

The total soil moisture in the  $n^{\text{th}}$  layer is given by:

$$\frac{dM_n}{dt} = W_{n-1} - W_n - E_n$$

(Equation 6.8)

Here, the total soil moisture in the layer is incremented by the diffusive water flux coming in from the layer above,  $W_{n-1}$ , the diffusive flux flowing out of the layer to the layer below,  $W_n$ , and the evapotranspiration extracted from the layer by plant roots and soil evaporation,  $E_n$ .  $E_n$  is calculated from the total evaporation based on the profiles of soil moisture and root density (Essery *et al*, 2001).

The diffusive vertical water flux for each layer,  $W$  ( $\text{kg m}^{-2} \text{s}^{-1}$ ), is given by the Darcy equation:

$$W = K \left( \frac{\delta\psi}{\delta z} + 1 \right)$$

(Equation 6.9)

where  $K$  ( $\text{kg m}^{-2} \text{s}^{-1}$ ) is the hydraulic conductivity and  $\psi$  (m) is the soil water suction.

These are determined by:

$$K = K_{sat} \left( \frac{\theta_u}{\theta_{sat}} \right)^{2b+3}$$

(Equation 6.10)

and

$$\psi = \psi_{sat} \left( \frac{\theta_u}{\theta_{sat}} \right)^{-b}$$

(Equation 6.11)

where  $K_{sat}$  is the saturated hydraulic conductivity,  $\psi_{sat}$  is the saturated soil water suction,  $\theta_u$  and  $\theta_{sat}$  are the unfrozen volumetric soil water and saturated volumetric soil water fractions and  $b$  is an empirical exponent.

The relationship between unfrozen soil water concentration and temperature when ice is present is described in terms of the soil water suction (Miller, 1965; Black and Tice, 1989 cited in Cox *et al*, 1999):

$$\psi = -k \left[ \frac{\rho_i L_f}{\rho_w T_m g} \right] (T - T_m)$$

(Equation 6.12)

where  $T_m$  (K) is the freezing point of pure water at atmospheric pressure,  $g$  ( $m s^{-2}$ ) is the acceleration due to gravity and  $k$  is a dimensionless constant that depends on the soil. The  $k$  constant is a measure of the degree to which the adsorption of the soil dominates over the capillarity and is used as a correction factor. This typically equals 1 for clay rich soils and 2.2 for granular soils. However, for organic soils, the value of the constant is uncertain.

The maximum volumetric fraction of unfrozen soil water that can exist at temperature  $T$ , where  $T < T_m$  is calculated by:

$$\frac{\theta_u^{\max}}{\theta_{sat}} = \left[ -k \frac{\rho_i L_f (T - T_m)}{\rho_w T_m g \psi_{sat}} \right]$$

(Equation 6.13)

The actual value of  $\theta_u$ , the unfrozen volumetric fraction, is limited by the total water content of the soil:

$$\theta_u = \min(\theta_u^{\max}, \theta)$$

(Equation 6.14)

where  $\theta$  is the “liquid” total volumetric concentration. This is the volumetric concentration that would exist if all of the soil water was in liquid form:

$$\theta = \theta_u + \frac{\rho_i}{\rho_w} \theta_f$$

(Equation 6.15)

For clarity, Equation 6.14 can be rewritten in terms of two distinct temperature regimes as follows:

$$\theta_u = \begin{cases} \theta_u^{\max} & \text{if } T < T_m \\ \theta & \text{if } T \geq T_m \end{cases}$$

(Equation 6.16)



where  $T_{\max}$  is the temperature above which all soil moisture is unfrozen.

In addition to the soil thermodynamics and soil water phase change processes, it is useful to note the boundary conditions for the MOSES scheme.

The top boundary condition is calculated in terms of the fluxes of throughfall precipitation ( $P_f$ ), snowmelt ( $S_m$ ) and surface runoff ( $Y_s$ ) as follows:

$$W_0 = P_f + S_m - Y_s$$

(Equation 6.17)

where  $W_0$  is the flux which infiltrates at the surface. Surface runoff occurs when the local water flux at the soil surface is greater than the saturated hydraulic conductivity (Cox *et al*, 1999).

The lower boundary condition is set to that of “free drainage” so that the drainage from the lowest soil layer ( $W_N$ ) equals the hydraulic conductivity of this layer ( $K_N$ ):

$$W_N = K_N$$

(Equation 6.18)

The equations and processes described above have been taken from Cox *et al*. (1999), Hall *et al*. (2003) and Essery *et al* (2001).

## 6.2 Modelling Framework

The MOSES land surface scheme was originally designed to be used in large climate models to represent the land surface. However, it has been increasingly used as a stand alone model for a number of hydrological investigations such as predicting wetlands and predicting river flows. This prompted the release of JULES, the Joint UK Land Environment Simulator. Based on MOSES, and having the same structure and land surface physics, JULES has been designed to be used and developed independently of the meteorological and climate models.

There were two main objectives of the modelling exercise. The first was to determine the optimum soil parameter values for the saturated hydraulic conductivity ( $K_s$ ), saturated soil water suction ( $\psi_s$ ) and the Clapp-Hornberger exponent ( $b$ ) to be used in the JULES Land Surface Scheme for the highly organic soils found in the study area using observed data. The second was to evaluate the performance of JULES for soil moisture and soil temperature and snow modelling during the seasonal freeze and thaw events observed at the study area. In order to achieve these objectives, a framework for the modelling exercise was created which can be seen in Table 6.1.

In order to run the model it was necessary to prepare two datasets. The first, a driving dataset, contained a continuous series of meteorological data collected at the fieldsite for the period to be modelled. These data included incoming shortwave radiation ( $\text{W m}^{-2}$ ), net longwave radiation ( $\text{W m}^{-2}$ ), rainfall rate ( $\text{kg m}^{-2} \text{s}^{-1}$ ), snowfall rate ( $\text{kg m}^{-2} \text{s}^{-1}$ ), surface air temperature (K), zonal wind speed ( $\text{m s}^{-1}$ ), meridional wind speed ( $\text{m s}^{-1}$ ), surface pressure (Pa) and specific humidity ( $\text{kg kg}^{-1}$ ). Observed rainfall and snowfall values (recorded in mm) were converted into rates ( $\text{kg m}^{-2} \text{s}^{-1}$ ) for use within the model. The driving data were half hourly averaged measurements (logged every minute) recorded by the automatic weather station on the Windfetch Grid (AWS1). Gaps in the driving data from AWS1 were filled by corresponding data from other weather stations at the fieldsite or from the Abisko research station. This driving dataset was prepared by Andrew Wiltshire (University of

Durham) and was then quality controlled (to remove any erroneous measurements) before being used for this modelling exercise.

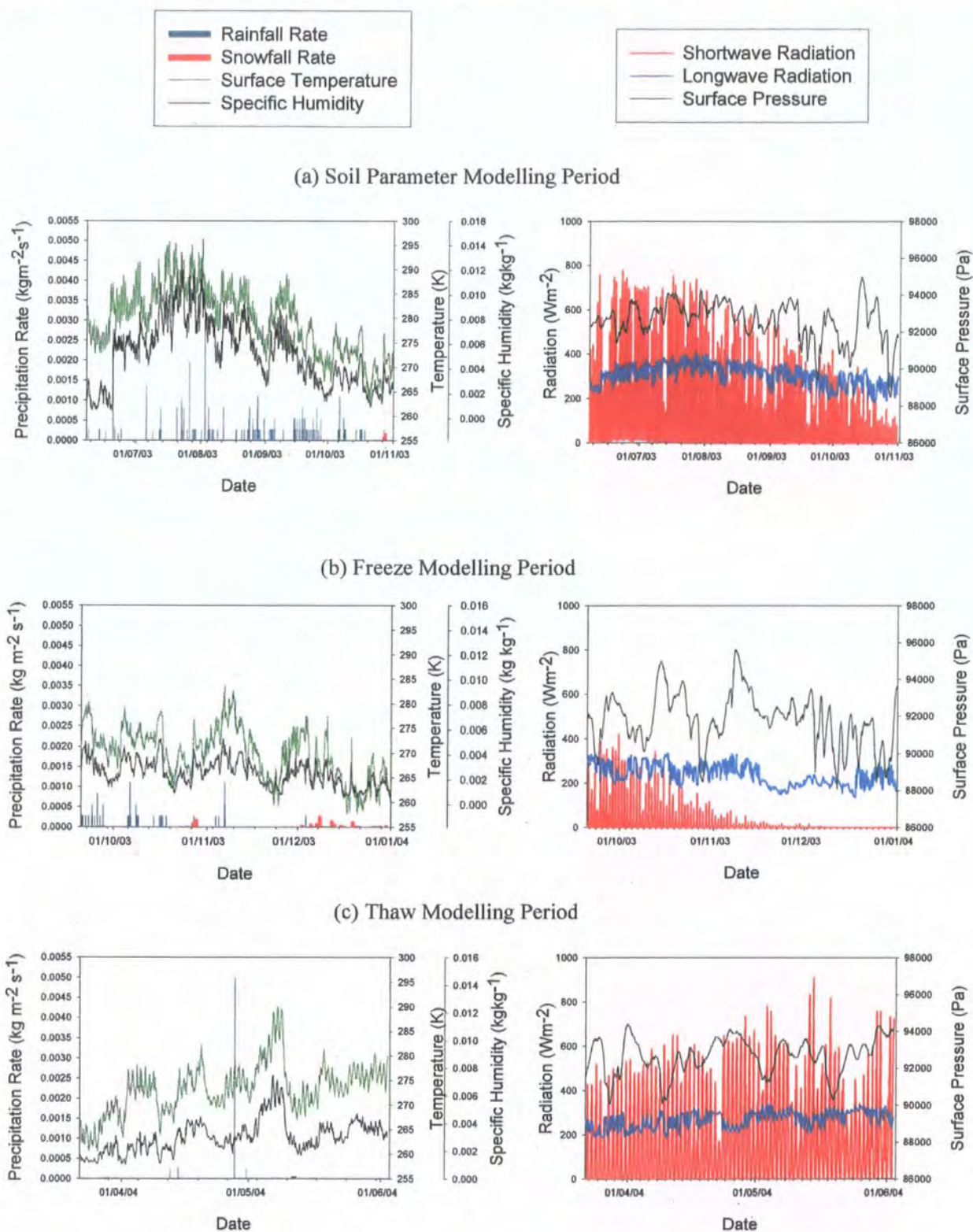
**Table 6.1 Modelling framework describing the objectives of the modelling exercise, description of the model runs and outputs of the model**

Objective	Model Runs	Output
Optimise soil parameter values for $K_s$ , $\psi_s$ and $b$ for organic soils.	1. 20 parameter sweep runs to determine model performance.	Time series of modelled soil moisture.
	2. 9000 parameter seep runs with random combinations of soil parameter values.	Modelled dry-down slope values against values of $K_s$ , $\psi_s$ and $b$ to be compared with observed slope values.
Evaluate model performance during the freeze and thaw periods.	1. JULES run to compare observed soil temperature with modelled soil temperature and snowmass for the freeze period.	Time series of modelled soil temperature in 1 <sup>st</sup> and 2 <sup>nd</sup> soil layers and gridbox snowmass to be compared with observed soil temperature.
	2. JULES run to compare observed soil temperature with modelled soil temperature and snowmass for the thaw period.	Time series of modelled soil temperature in 1 <sup>st</sup> and 2 <sup>nd</sup> soil layers and gridbox snowmass to be compared with observed soil temperature.
	3. JULES run to compare observed soil moisture with modelled frozen and unfrozen soil moisture fractions and modelled snowmass.	Time series of modelled frozen and unfrozen soil moisture fraction and gridbox snowmass to be compared with observed soil moisture.

Figure 6.1 shows the driving dataset used for the three modelling periods. The summer of 2003 was chosen for the soil parameter modelling period, starting on the 7<sup>th</sup> June 2003 and ending on the 1<sup>st</sup> November 2003. This ensured that the soil was not frozen during the soil parameter sensitivity modelling which would have complicated the analysis. The freeze modelling period was determined by the date that the observed soil temperature dropped below zero (at 5 cm depth) and was between the 20<sup>th</sup> September 2003 and the 1<sup>st</sup> January 2004. The thaw period, when the observed soil temperature rose above zero, started on the 22<sup>nd</sup> March 2004 and ran until the 2<sup>nd</sup> June 2004.

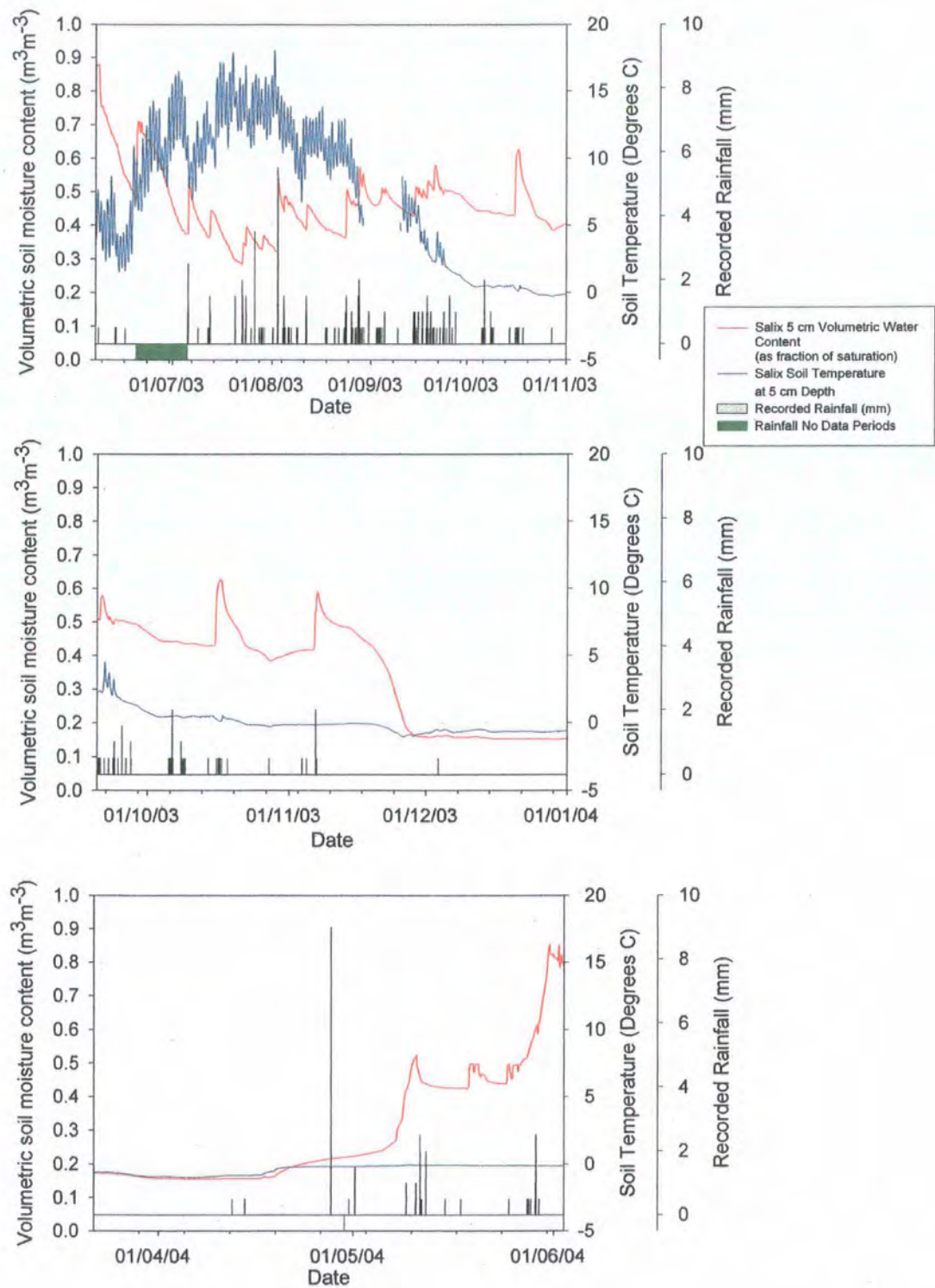
The second dataset is used to validate that the processes are being correctly represented and contain observed soil moisture and soil temperature at 5cm depth and recorded rainfall. The soil moisture and soil temperature data were taken from the hillslope logger at the *Salix* site, in a topographic hollow. These data were used, in preference to the *Empetrum* site, since any effect of laying snow on the modelled soil moisture and temperature regime would be more pronounced here (since the wind blown snow accumulates in the topographic hollow) and it would give an insight into the model performance in the presence of laying snow. The *Salix* site also provided a continuous observed dataset of soil moisture and temperature for the modelling periods. The data were taken at 5 cm depth since the observed data could then be compared with the top soil layer of the JULES soil component (an average taken over the top 0.1 m). The rainfall data were taken from the Windfetch Grid automatic weather station (AWS1). All the observed data were recorded as half hourly averages.

The observed data used for the three modelling periods can be seen in Figure 6.2. The graphs also show periods when there were no observed rainfall data available due to technical errors in the measurement process.



**Figure 6.1 Driving Data used in the JULES Land Surface Scheme for the (a) Soil Parameter, (b) Freeze and (c) Thaw Modelling Periods**





**Figure 6.2 Observed Soil Moisture, Temperature and Rainfall at the Hillslope *Salix* Site for the (a) Soil Parameter, (b) Freeze and (c) Thaw Modelling Periods**



6.3 Soil Parameter Sensitivity

Currently, soil parameter values for three soil types (clay, loam and sand) can be obtained from a look-up table for use within MOSES. A soil type is chosen for the model run and the parameter values are used for all of the soil layers. JULES, however, allows different soil types to be used for each soil layer but currently the default values are those used in MOSES. Since the soils of the study area are predominantly organic (refer to Chapter 5), the current soil parameter values are not representative. Lloyd *et al.* (1999) (cited in Hall, unpublished), suggest parameter values for an organic soil that appear to be more appropriate: The current soil parameter values and those suggested for an organic soil are presented in Table 6.2.

**Table 6.2 Soil parameter values currently used in MOSES/JULES and those suggested for organic soils (Lloyd *et al.*, 1999, cited in Hall, unpublished)**

Parameter	Description	Typical Value			
		Clay	Loam	Loamy Sand	Hall (Organic)
b	Clapp-Hornberger exponent	11.2	6.6	3.6	4
C <sub>s</sub>	Soil Heat Capacity (J m <sup>3</sup> K <sup>-1</sup> )	1.23E+06	1.19E+06	1.32E+06	1.00E+06
λ	Soil Thermal Conductivity (W m <sup>-1</sup> K <sup>-1</sup> )	0.22	0.23	0.32	0.025
K <sub>s</sub>	Saturated Hydraulic Conductivity (kg m <sup>-2</sup> s <sup>-1</sup> )	0.0036	0.0047	0.011	0.015
ψ <sub>s</sub>	Saturated Soil Water Suction (m)	0.045	0.049	0.022	0.015
θ <sub>s</sub>	Volumetric Soil Moisture at Saturation (m <sup>3</sup> m <sup>-3</sup> )	0.456	0.458	0.382	0.8
θ <sub>c</sub>	Volumetric Soil Moisture at Critical Point (m <sup>3</sup> m <sup>-3</sup> )	0.31	0.242	0.096	0.3
θ <sub>w</sub>	Volumetric Soil Moisture at Wilting Point (m <sup>3</sup> m <sup>-3</sup> )	0.221	0.136	0.033	0.1

There are, however, many classes of organic soil, each with varying properties. Letts *et al.* (2000) present a new parameterisation for organic soils for use in the Canadian Land Surface Scheme (CLASS). Particular attention is paid to the hydraulic conductivity of different peat soils (Fibric, Hemic and Sapric) and a number of observed values are presented from the literature. A selection of these values can be seen in Table 6.3.

This small selection taken from Letts *et al.* (2000) shows that the range of hydraulic conductivity values can vary from  $2.2 \times 10^{-3} \text{ (ms}^{-1}\text{)}$  (Fibric peat) to  $1 \times 10^{-12} \text{ (ms}^{-1}\text{)}$  (Sapric peat). Bellisario *et al.* (2000) suggest values of hydraulic conductivity for use in the CLASS for each peat type ( $1.7 \times 10^{-4}$  Fibric,  $2.0 \times 10^{-6}$  Hemic and  $1.0 \times 10^{-7}$  Sapric). Also suggested in this paper are values of the Clapp-Hornburger exponent (the  $b$  exponent in Equations 6.6 and 6.7) ranging from 2.7 for Fibric peat to 12.0 for Sapric peat.

Modelling highly organic peat soils can be extremely complex due to their special properties. For example, the hydraulic conductivity of peat soils typically varies with depth. Also, the thermal conductivity of these soils varies considerably depending on their wetness state (dry or saturated) and whether they are frozen or unfrozen.

However, for the purposes of this modelling exercise, it was assumed that the hydraulic properties of the soil were uniform with depth and the model's default value for the thermal conductivity an organic soil was used. Also, the model assumes that the soil properties do not show seasonal variations. Future work could address these issues (as discussed in Section 7.5.6).

Throughout the modelling exercise, the model default parameter values for snow and soil albedo and thermal conductivity were used. The model uses different values for albedo depending on the snow state; if snow is not present, a value of -1.0 is used for soil but if snow is present, a value of 0.8 is used. Similarly, the thermal conductivity of snow ( $0.265 \text{ W m}^{-1} \text{ K}^{-1}$ ) and soil ( $0.025 \text{ W m}^{-1} \text{ K}^{-1}$ ) are used depending on the snow state.

**Table 6.3 Hydraulic conductivity observations (after Letts *et al.*, 2000)**

<b>K (m/s)</b>	<b>Peat Quality Observations</b>	<b>Source</b>
<b>Fibric peat</b>		
$2.2 \times 10^{-3}$	von Post humification index = 1	Gafni (1986)
$1.7 \times 10^{-4}$	von Post humification index = 3	Maelstrom, 1923 in Ingram <i>et al.</i> (1975)
$5 \times 10^{-5}$	Slightly decomposed fen peat	Romanov (1968)
$3 \times 10^{-6}$	von Post humification index = 1.5	Ryden(1990) in Magnussen (1994)
$3 \times 10^{-7}$	von Post humification index = 3	Ryden(1990) in Magnussen (1994)
<b>Sapric peat</b>		
$5 \times 10^{-4}$	von Post humification index = 4	Gafni (1986)
$8 \times 10^{-6}$	Moderately decomposed fen peat	Ivanov (1981)
$6 \times 10^{-7}$	von Post humification index = 3.5	Ryden(1990) in Magnussen (1994)
$6 \times 10^{-8}$	von Post humification index = 6	Ryden(1990) in Magnussen (1994)
$1 \times 10^{-9}$	von Post humification index = 7	Ryden(1990) in Magnussen (1994)
<b>Hemic peat</b>		
$5 \times 10^{-6}$	<i>Carex</i> peat, von Post humification index = 8	Sarasto (1961) in Ingram <i>et al.</i> (1975)
$5 \times 10^{-7}$	Highly decomposed fen peat	Romanov (1968)
$2 \times 10^{-7}$	Humic subsurface peat	Boelter (1968)
$2 \times 10^{-10}$	von Post humification index = 8	Ryden(1990) in Magnussen (1994)
$1 \times 10^{-12}$	von Post humification index = 9	Ryden(1990) in Magnussen (1994)

The Monte-Carlo parameter sweep, developed by Dr. Eleanor Blyth and Dr. Douglas Clark at CEH Wallingford (unpublished, pers. comm.) was designed for the calibration of the JULES model and can be used to explore the sensitivity of JULES to the model parameters. The parameter sweep allows certain parameter values in JULES to be varied between user-defined limits so that calibration and sensitivity analysis can be carried out. The values are then randomly chosen between the lower and upper limits. The parameter sweep routine carries out a specified number of JULES runs with the different combinations of parameter values. A FORTRAN script is then used to perform statistical analysis of the output data before a final output file is created.

An initial parameter sweep was carried out varying values for the  $b$  exponent, saturated soil water suction ( $\psi_s$ ), saturated hydraulic conductivity ( $K_s$ ), soil heat capacity ( $C_s$ ) and soil thermal conductivity ( $\lambda$ ). The lower and upper parameter values chosen for this analysis for the parameter sweep are shown in Table 6.4. These were chosen to span the values currently used in JULES and those suggested in the literature.

Initially, 20 runs of the JULES model were carried out for the soil parameter modelling period (7<sup>th</sup> June to 1<sup>st</sup> November, 2003). The output from this initial sweep was the unfrozen volumetric soil moisture fraction in the top soil layer of the model. The model output was compared to the observed values of volumetric soil moisture to determine the overall performance of the model before choosing a way to quantify that performance.

**Table 6.4 Initial parameter sweep upper and lower value limits**

Parameter	Lower Limit	Upper Limit
$b$	1	12
$\psi_s$	0.02	0.05
$K_s$	0.0001	0.1
$C_s$	6.00E+05	2.00E+06
$\lambda$	0.02	0.2

Figure 6.3 shows the output from these initial model runs. The modelled soil moisture was underestimated for most of the model runs. The input data were amended to include a 15 mm rainfall event on the 19<sup>th</sup> June 2003 (labelled A in Figure 6.3) since the observed soil moisture suggested an influx of water at this time and the existing driving dataset did not include a rainfall event (due to the absence of recorded rainfall data due to technical problems). Also, after the 18<sup>th</sup> September 2003 (labelled B in Figure 6.3), the model does not perform well, probably due to snowfall events which were recorded at this time.

By studying this figure, it was clear that the overall soil moisture was sensitive to the dry-down after a rainfall event and that the hydraulic performance of the model during these drying events dominates the movement of water through the soil. In these model runs with default parameters, the model runs consistently exaggerated the drying of the soil. The steeper slopes displayed by the model runs compared to the observed data indicate a greater degree of drying; the soil water apparently flowing laterally through the top soil layer of the model faster than is actually observed.

It should be noted that there are alternative explanations for the flattening of the slope during these drying events. It is possible that the saturated hydraulic conductivity of the soils decrease with depth (as is often the case with these types of soil). This would cause the soil moisture storage to be filled and increase the soil moisture during the study period. Also, the lateral movement of water into the soils would again increase the overall soil moisture and result in the flattening of the observed slopes during the drying events. However, since JULES is a one dimensional model and can only simulate the vertical fluxes of soil water, the lateral movement of water can not be included. The calibration may therefore produce 'effective' parameters which describe the water balance as though the land were homogeneous and flat and the soil parameters are uniform with depth.

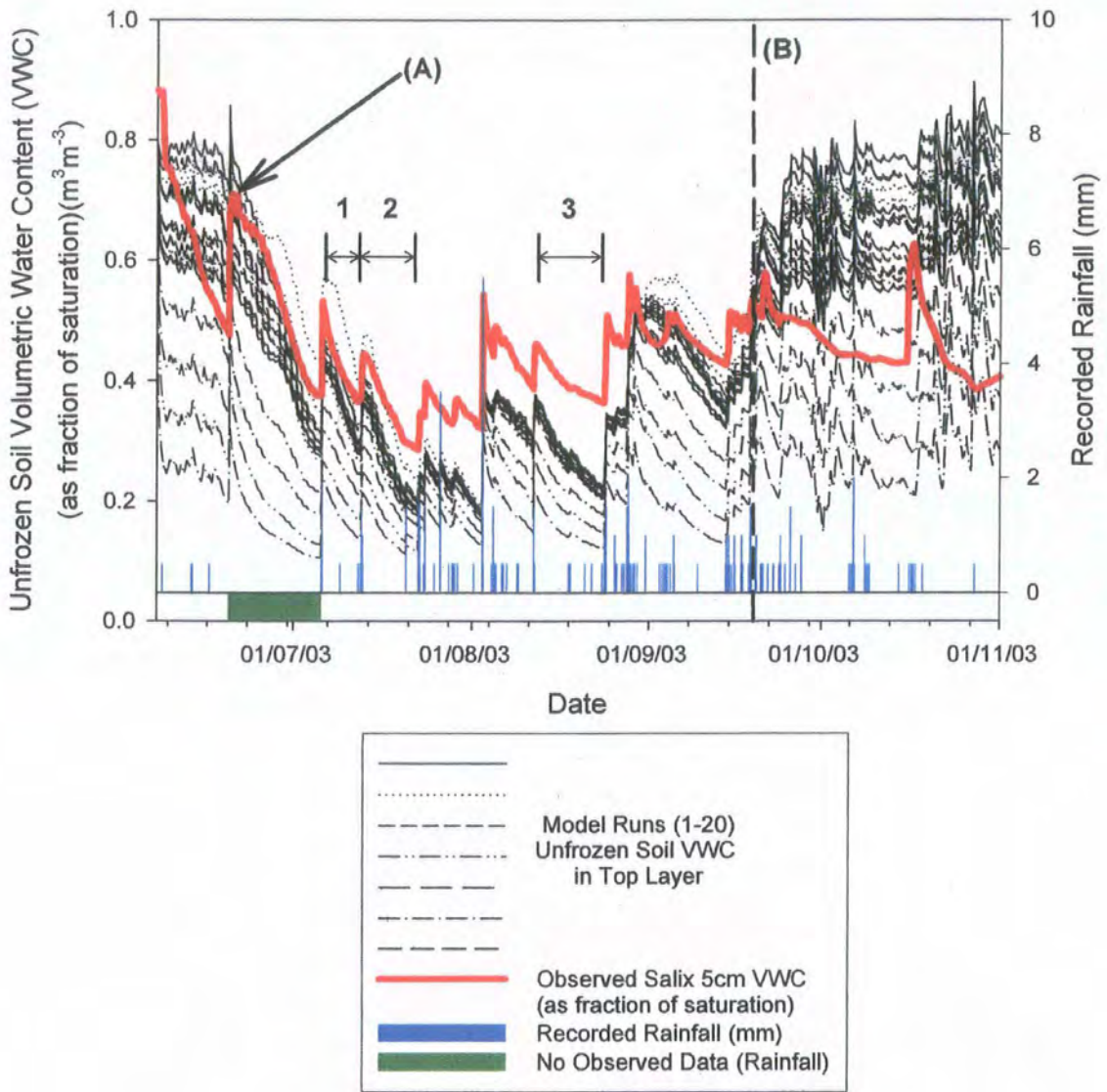


Figure 6.3 Monte-Carlo parameter sweep output varying  $b$ ,  $\psi_s$ ,  $K_s$ ,  $C_s$  and  $\lambda$



It was decided to perform a calibration using three soil drying slopes (labelled 1, 2 and 3 in Figure 6.3). These slopes were determined by the change in soil moisture during the dry-down events:

$$\text{Drying Slope} = \frac{\theta_w - \theta_d}{\text{time}}$$

(Equation 6.19)

where  $\theta_w$  and  $\theta_d$  are the unfrozen volumetric soil water (as a fraction of saturation) at the start and end (respectively) of the drying event.

The model was run 9000 times, with random combinations of the soil parameters using the same lower and upper limits given in Table 6.4. The random selection of  $K_s$  values was taken from the log  $K_s$  values. The observed and modelled drying slopes for each of the identified periods were calculated.

Figures 6.4, 6.5 and 6.6 show the modelled slope (y axis) values against the values of saturated hydraulic conductivity, b exponential and soil water suction (respectively) (x axis) for each of the slopes. The observed slope values are shown with a +/- 2.5% error region which represents the measurement error of the CS616 Water Content Reflectometer (Campbell Scientific Inc.) used to take the measurements in the field.

Figure 6.4 shows the slope against the saturated hydraulic conductivity. For all three slopes, the modelled slope values are sensitive to the hydraulic conductivity. The optimum modelled slope occurs between 0.0001 and 0.001.

The model outputs show that the results are also sensitive to the b value (Figure 6.5). Here, the optimum value is between 1 and 4 (slope 1) and at around 4 for slope 2. The outputs from Slope 3 show less sensitivity and no optimum value can be taken from these runs. On the basis of the modelled output for slopes 1 and 2, the optimum value for the b exponential should be 4.

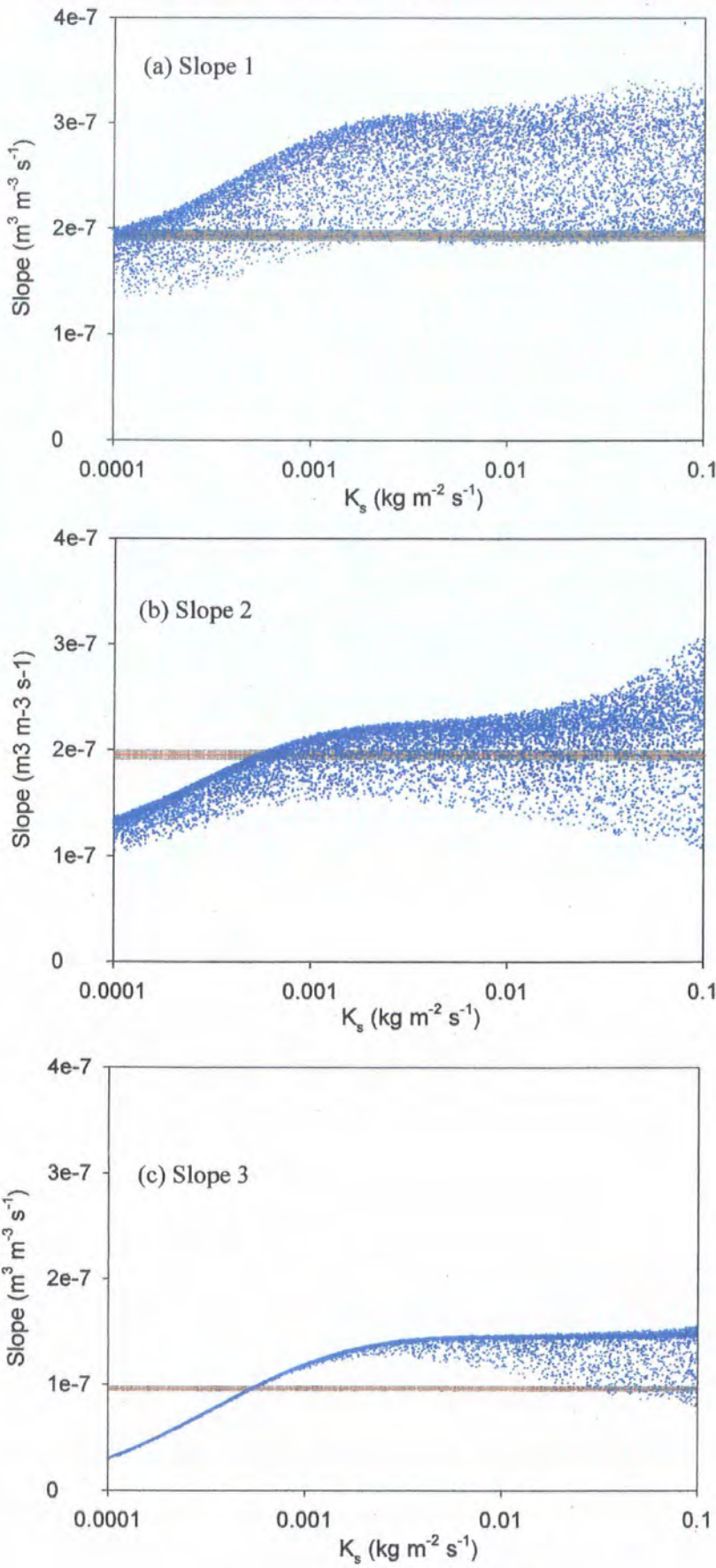
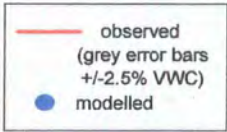


Figure 6.4 Graphs showing the results of Monte Carlo analysis for varying values of the  $K_s$  (saturated hydraulic conductivity) parameter for three soil moisture dry-down events (slopes) during the Summer 2003.



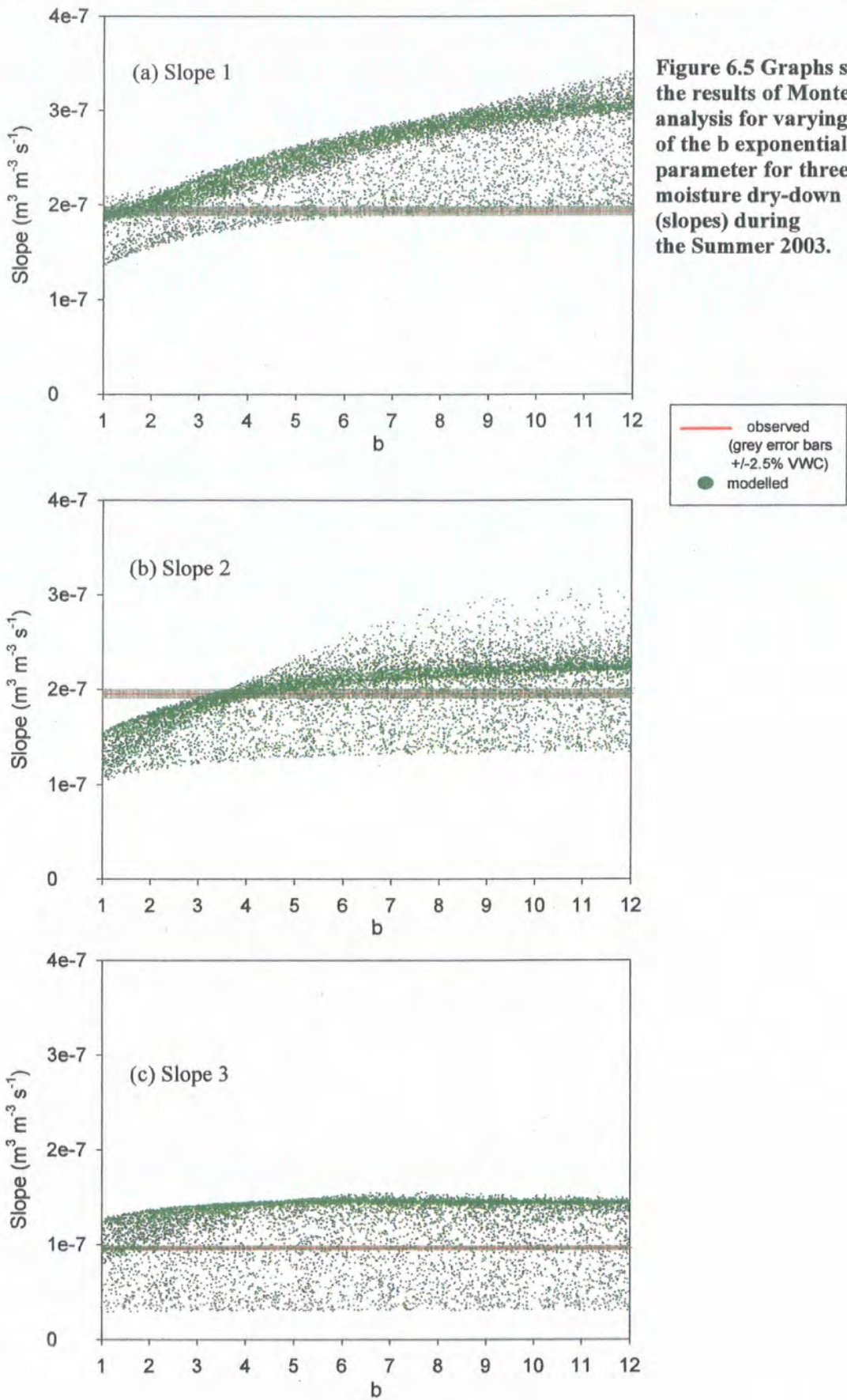


Figure 6.5 Graphs showing the results of Monte Carlo analysis for varying values of the  $b$  exponential parameter for three soil moisture dry-down events (slopes) during the Summer 2003.



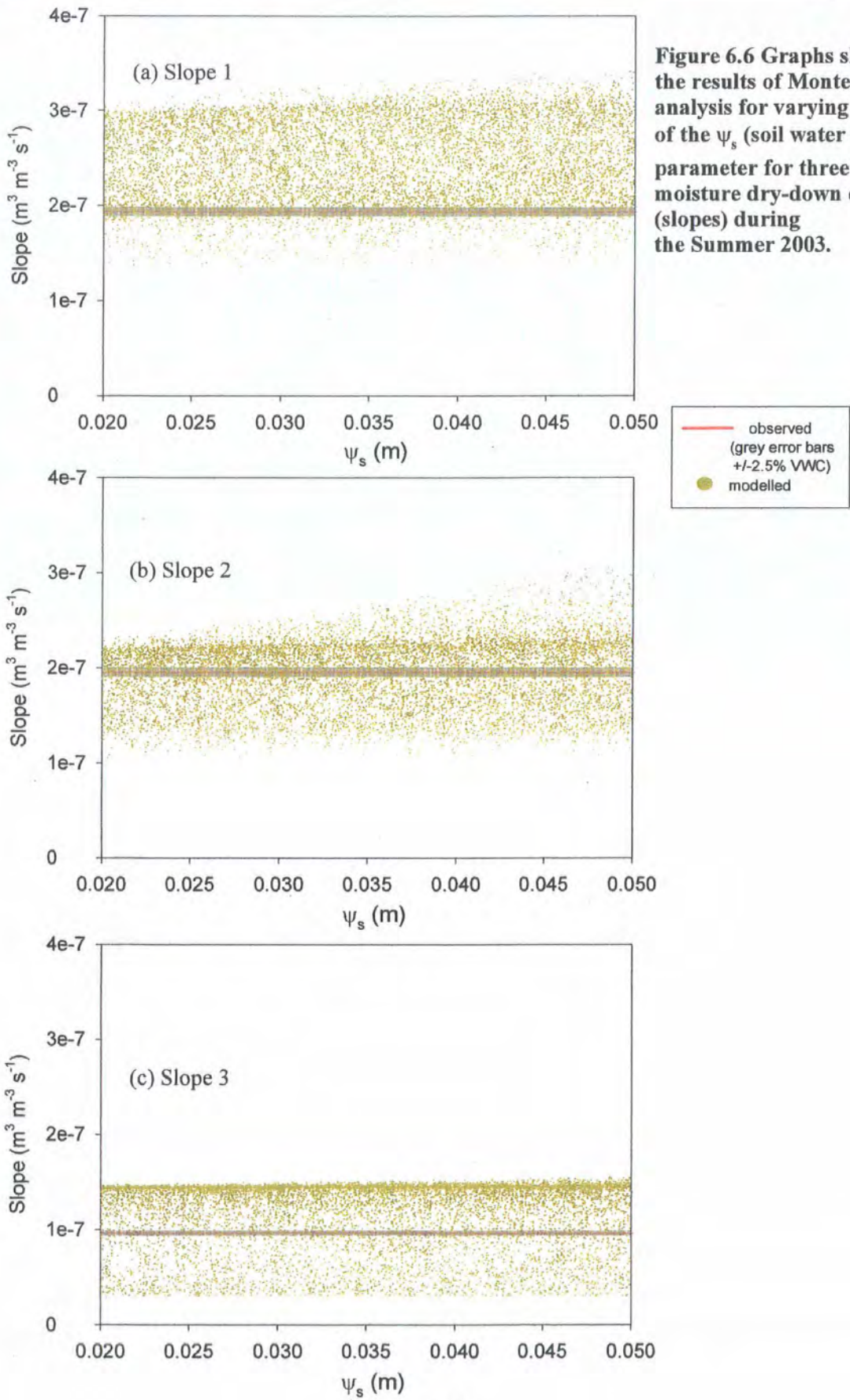


Figure 6.6 Graphs showing the results of Monte Carlo analysis for varying values of the  $\psi_s$  (soil water suction) parameter for three soil moisture dry-down events (slopes) during the Summer 2003.

Figure 6.6 shows the slopes with respect to the soil water suction. It is clear that the drainage is not sensitive to this parameter, all points being evenly distributed across the chosen parameter value limits.

The methods outlined above have shown how optimum model soil parameters can be determined from observed data. The results show that, while the  $b$  exponential derived by this method is the same as that proposed by Hall (unpublished) and Lloyd *et al.* (1999), the saturated hydraulic conductivity is an order of magnitude lower than suggested.

#### 6.4 Model Performance During Freeze/Thaw Events

The performance of the JULES model during the soil freezing and thawing periods has two main aspects which are linked. The first is the heat flux into and out of the soil. The second is the moisture flux within the soil which, during these periods, is complicated by the presence of both unfrozen and frozen soil water.

The JULES model was run for both a freeze and a thaw period, using the soil parameter values obtained from the parameter sweep analysis, and the output was compared with observed values of soil temperature and soil moisture so that the performance of the model could be assessed.

##### 6.4.1 The Freeze Period

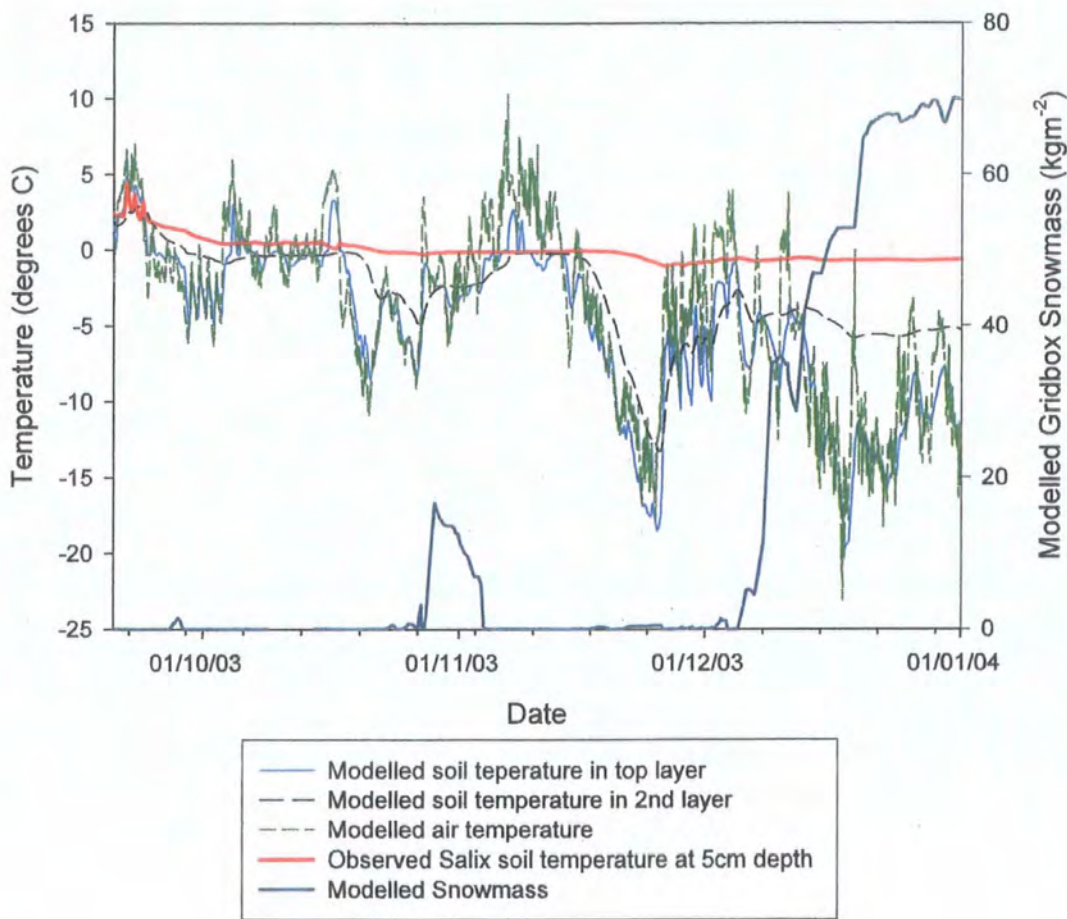
During the freeze period, the main concern was that the model was not simulating the soil temperature with any accuracy which is due, in part, to the inability of JULES to represent correctly the presence of snow cover (especially spatially distributed snow cover due to blowing snow). The snow cover has a big impact on soil temperature because the snow acts to insulate the underlying soil from the very variable air temperatures. This is demonstrated in the observations (Figure 6.7) where the observed soil temperature is seen to follow the air temperature closely for the first few days and then to gradually fall to below zero and remain around zero for the whole period. Although we have no direct measurements of

snow cover at this particular site observations made in the field suggest that the snow cover can remain on the landscape after a snowfall event even though the air temperature subsequently rises above zero. Figure 6.8 shows the location of the observed soil temperature used in the soil freezing model runs on the 10<sup>th</sup> November 2003. The snow cover here persisted in the topographic hollows.

A JULES run was carried out to compare the observed soil temperature at 5 cm depth with the modelled soil temperature in the top two soil layers. Figure 6.7 shows the model output for the freeze period. Also displayed is the modelled gridbox snowmass which represents the amount of snow cover. The most striking result of this model run is that the modelled soil temperature in the top soil layer follows the air temperature very closely. The modelled top layer temperature rises (to a maximum of around 10 °C) and falls (to a minimum of around -23 °C) in unison with the air temperature. During periods of snow cover, this top layer is within the snowpack and is not intended to be representative of the soil conditions below. Instead, the second soil layer temperature is much closer to the observed values at the start of the modelling period and appears much more stable until the middle of November when the modelled snowpack disappears and the insulation effect is gone; dropping to around -15 °C before rising and levelling out at around -5 °C.

Three periods of modelled snow cover can be distinguished in Figure 6.7. A short period occurs at the end of September; followed by another in late October with greater snowmass present. A longer period follows in early December with a large amount of snow present. In this last period, the insulating effect of the snow on the soil temperature can be seen. However, the temperature in the model is much lower than the observed. This may be due to the observed soil temperatures being affected by the input of water due to lateral seepage which would act to maintain soil temperatures at around zero during the freezing period. The model would not pick up on this since it is one-dimensional and only deals with vertical fluxes of water. This could explain the modelled soil temperatures being considerably lower than the observed during this period.





**Figure 6.7 Soil Temperatures (modelled and observed), Air Temperature (modelled) and Laying Snow (modelled) for the Freeze Modelling Period at the *Salix* site**



**Figure 6.8 Photograph showing the *Salix* site on the Hillslope Grid during the freeze modelling period. The snow is seen to persist in the topographic hollow  
(Photograph taken on 10<sup>th</sup> November 2003)**

The observed soil temperatures are insulated, to some degree, by the snowpack. This would also account for the observed temperatures being higher and more stable than the air temperatures.

#### 6.4.2 The Thaw Period

During the thaw period, the modelling of soil moisture is complicated by the presence of both the frozen and unfrozen fraction and also the behaviour of the soil under the snowpack. JULES model runs were carried out to investigate both the modelled soil temperature and soil moisture for the thaw period of 2004. The model was initialised with the amount of soil moisture present before the soil froze and also  $75 \text{ kg m}^{-2}$  of laying snow, which is representative of the conditions observed at the fieldsite at this time.

Figure 6.9 shows the modelled temperature of all four soil layers during the thaw period. Layers 3 and 4 are most similar to the observed soil temperature, being more stable than the top two layers. This is because, when snow is present, the JULES model treats the top two soil temperature layers as a snow/soil mix. The temperature in the top layer is, therefore, effectively the snowpack temperature (depending on the snow depth).

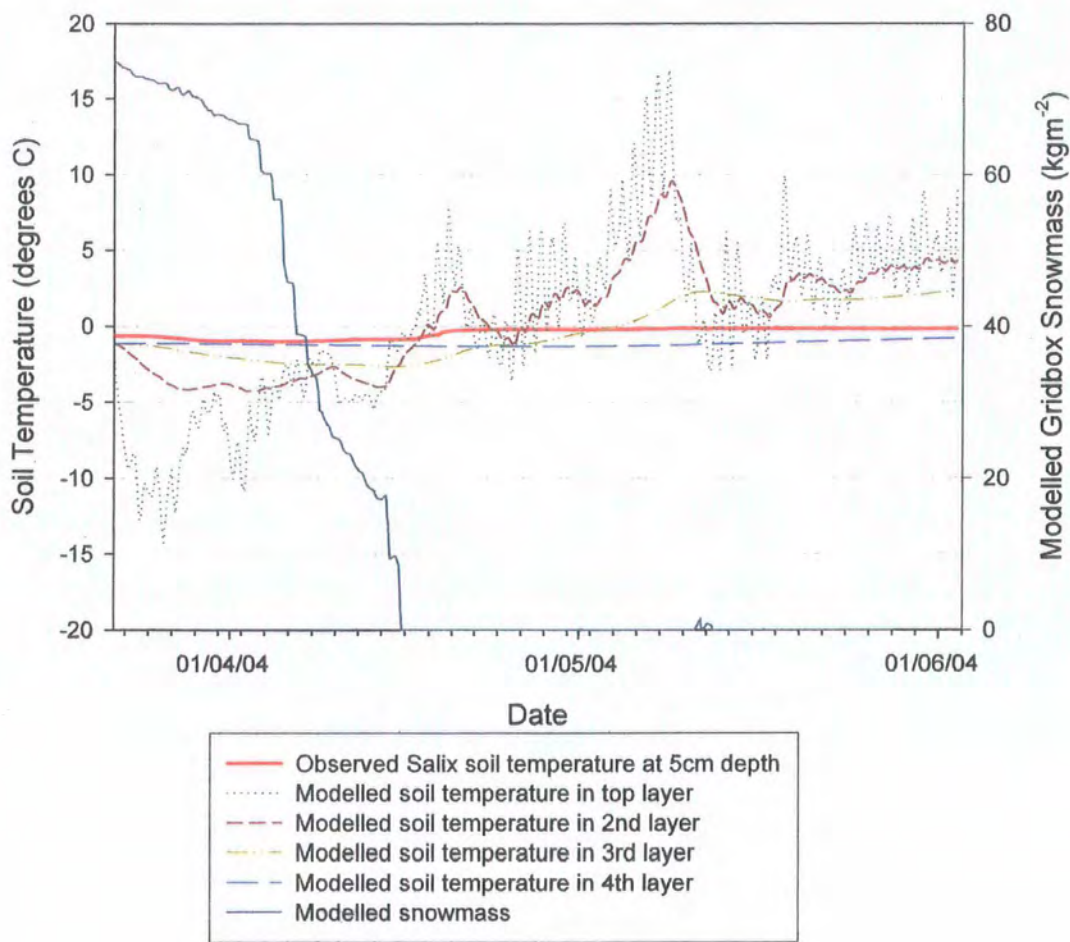


Figure 6.9 Soil Temperatures (modelled and observed) and Laying Snow (modelled) for the Thaw Modelling Period at the *Salix* site



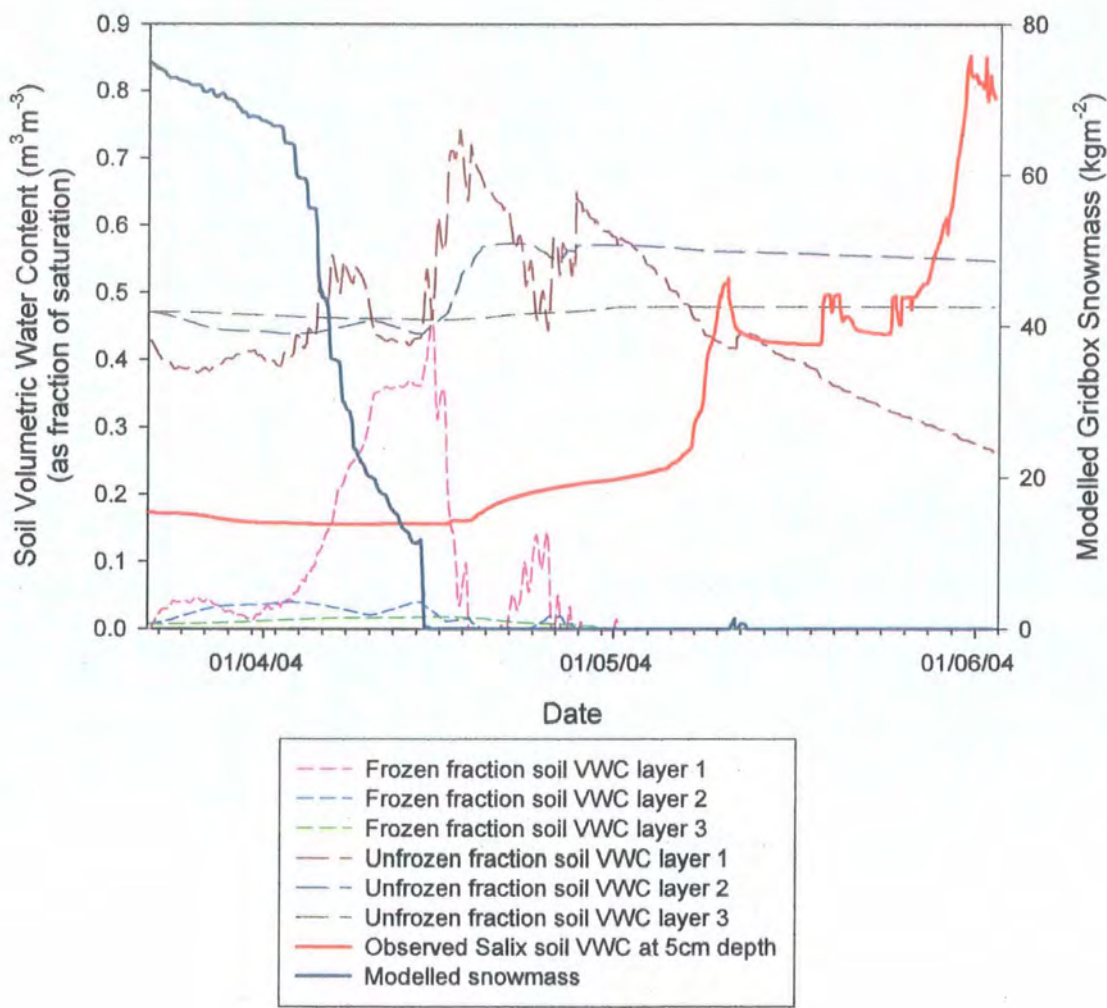


Figure 6.10 Soil Volumetric Water Content and Laying Snow for the Thaw Modelling Period at the *Salix* site

The snow cover is seen to rapidly decrease as the modelled temperatures in the top two layers rise above zero. However, this is unrealistic when compared to observed conditions at the fieldsite since the snow cover has been observed at the fieldsite until late May or June when air temperatures are above freezing. The presence of snow cover in reality would explain the observed soil temperature being stable at around zero for the whole thaw modelling period. Also, the lateral movement of water would influence the observed soil temperature, maintaining it at around zero during the thaw period. Again, the model would not take this into account due to it being one-dimensional.

Figure 6.10 shows both the modelled frozen and unfrozen fractions of soil volumetric water content for three soil layers during the thaw period. The frozen fraction in the top soil layer is present until mid to late April (when the soil temperature in this layer begins to rise above zero – see Figure 6.9). In layers two and three, the frozen fractions are small at the beginning of the period and reduce to zero in late April. The unfrozen fraction in the top layer peaks at around  $0.75 \text{ m}^3 \text{ m}^{-3}$  in mid April when the modelled snow cover reduces to zero. This indicates that the modelled snowmelt water is infiltrating straight into the soil at this point. It should be noted that the unfrozen fractions of all soil layers bear no resemblance to the observed soil moisture pattern. This is due to the continued presence of snow cover and frozen soils observed at the fieldsite until mid-May, and therefore the mismatch between model and observation conditions is quite severe.

### 6.5 Evaluation of Model Performance

The modelling exercise outlined in this chapter has highlighted a number of issues that should be addressed when using the model for high latitude studies. These include the ability of the model to simulate soil temperature in the presence of snow cover and soil water freezing and thawing regimes.

When modelling freeze and thaw events, it appears that direct comparisons with observed soil temperature and soil moisture are not useful when attempting to understand the underlying processes involved at the *Salix* site. This is mainly due to the fact that a one

dimensional model can not predict the overlying snow conditions at a particular point in a wind-swept landscape where snow drifts are common and deep.

At present, the modelled snow cover thaws quickly when the air temperature rises above zero. In practice this is not the case as the snow persists on the landscape for a long time after the air temperature rises above freezing. This has implications for the soil temperature and soil moisture. The modelled soil temperatures are modified in the presence of snow so that the top two layers are effectively a snow/soil mix that is then strongly influenced by the air temperature. The temperature at the surface of the snowpack is taken to be the temperature at the surface of the top soil layer. The effect is that the third soil layer is much more representative of the actual measured soil temperatures at shallow depth. The insulation effect of the snowpack appears to be much more pronounced (from the measured soil temperatures) than that modelled.

The insulation of the snow could be modified by changing the snow thermal conductivity and the roughness of the land surface which affects the heat exchange with the atmosphere.

In a partially frozen soil, the representation of soil water in the model is inextricably linked to the soil temperature. The presence of a frozen fraction complicates the modelling of the soil water and causes changes in the hydraulic conductivity and soil water suction. Also, the snowmelt from the modelled snowpack tends to infiltrate the soil quickly. This, coupled with the early melting of snow cover in the model (compared to field observations), causes the unfrozen soil water fraction to peak just after the snowmelt. Observations made in the field suggest that the snowmelt from a snowpack occurs as overland flow, the soil still being frozen at the base of the snowpack. This water is then redistributed and accumulates to form localised ponding and wet areas (refer to Figure 3.11 May and June 2003 and May 2004).

The modelling of the freeze and thaw periods carried out in this chapter used observations of soil temperature and moisture at the *Salix* site to evaluate model performance. However, it is apparent that it would be difficult to simulate the soil moisture and temperature



regimes at this location accurately given the local effects of wind-blown snow and lateral water movement that are associated with the position of the site on the landscape. The observed values of soil temperature and moisture would have been affected by the lateral movement of water. This influx of water during the freeze and thaw periods, when spatially distributed patches of frozen soil exist, would undergo a phase change and act to maintain soil temperatures at around zero during these periods. This is shown by the observed soil temperatures at the *Salix* site. However, the model does not consider a lateral influx of water (being one-dimensional the model only deals with vertical fluxes), and the simulated soil temperatures vary considerably from the observed. When the modelled soil temperatures (Figures 6.7 and 6.9) are compared with those observed at the Windfetch logger site (Figures 3.18 and 3.19) for the freeze and thaw periods, the model appears to have performed better. This may be due to the Windfetch logger site being located on a site less prone to lateral water movement (flatter topography) and snowdrift accumulation.

Therefore, although the model does not perform well at the *Salix* site, the results of the modelling of the freeze and thaw events were better at the Windfetch site. This suggests that it may be the local effects of increased snow and lateral water movement experienced at the *Salix* site that causes discrepancies between the modelled and observed soil temperature and moisture.

## 6.6 Chapter summary

Modelling exercises have been carried out using the JULES land surface scheme in order to determine model parameter values using a new Monte-Carlo parameter sweep method and also to assess the performance of the scheme during periods of soil freezing and thawing.

The parameter sweep method made it possible to perform a large number of model runs with different combinations of soil parameter values, the output of which was then analysed to diagnose the most appropriate values from observed soil moisture data. From this exercise it is apparent that the saturated hydraulic conductivity of the soil is the main factor controlling the movement of water through the soil. The parameter sweep results for the  $b$  exponent (used to calculate the soil water suction and hydraulic conductivity) agreed with the value for an organic soil recommended in the literature. The saturated hydraulic conductivity, however, was much lower than the suggested value.

During the period of soil freezing, the modelled soil temperatures in the top two soil layers did not correlate with the observed soil temperature at 5 cm depth at the *Salix* site. In the top soil layer, the soil temperature closely followed the air temperature. The modelled snow cover disappeared when the air temperature rose above zero whilst field observations suggest that the snow persists in drifts even though the air temperature is above freezing.

The main issue to arise from the thaw period was the infiltration of the snowmelt straight into the soil. From observations made in the field, this appears not to be the case. The meltwater occurs as overland flow (over frozen soils) and is redistributed across the landscape causing localised ponding. This is best illustrated in Figure 3.11, where the localised areas of highest soil moisture occur downslope of the snow patches and frozen soils observed in May 2003, June 2003 and May 2004.

During both the freeze and thaw periods, modelled soil moisture and soil temperature do not take into account the lateral redistribution of water due to topography. This lateral seepage would have a significant effect on observed soil moisture and temperature.

**Chapter 7**

**On the Spatial and Temporal Distribution of Near Surface Soil Moisture Across a  
Low-Arctic Tundra Hillslope**

**Discussion**

The work presented in this thesis has described the spatial and temporal measurement of near surface soil moisture across a heterogeneous, low Arctic landscape. Soil moisture was measured spatially over two contrasting grids, one on a hillslope (termed the 'Hillslope Grid') and the other a flatter site (named the 'Windfetch Grid'), during surveys in May, June, July, August and September 2003 and May, July, August and September 2004. Fine resolution time series data were also collected, giving an almost continuous record of near surface soil moisture and soil temperature for this period. High resolution digital terrain mapping data were obtained using manual surveying techniques and remote sensing which allowed the creation of a digital representation of the study area. Spatial surveys of soil depth, vegetation height and snow cover were also undertaken. The performance of the JULES Land Surface Scheme for high-latitude soil moisture and temperature studies has been evaluated.

This chapter will discuss the analysis of the data collected in this thesis and relates these findings to current thinking in this field. Conclusions arising from this study will be presented along with recommendations to take the work further.

### 7.1. Spatial and Temporal Soil Moisture Variability

A key aspect of this project was the description of the spatial and temporal variability of soil moisture across the Tundra landscape. This was necessary so that the underlying physical processes operating at the hillslope scale could be determined. To examine this fully, the analysis of data collected during the field campaigns is discussed in order to determine the hydrological processes operating at the study site.

The analysis of the spatial and temporal soil moisture data, collected during the field campaigns, has been presented in Chapter 3. The repeated measurement of soil moisture spatially across both the Hillslope site (180 m x 60 m) and Windfetch site (100 m x 200 m) at the 10 m scale provided data that were analysed using both standard statistical methods and geostatistics. Both the Hillslope and Windfetch sites displayed a large range in volumetric water content during the study period. The highest soil moistures were recorded

during the spring months (May and June) of both 2003 and 2004 across both sites. This is to be expected since it is during this period that the snow accumulated across the landscape over the winter months begins to thaw and the main water input into the hydrological system at this time is due to snowmelt.

Histograms produced using the spatially collected data (Figures 3.2 and 3.3) confirm that seasonality is displayed in the near surface soil moisture values for the two sites. In May of both years, the histograms are symmetrical and become more positively skewed as the summer season progresses, with higher frequencies towards the dry end of the distributions becoming more pronounced with time, from July to September. This shows that the initial wet conditions during the spring snowmelt in May and June give way to much lower soil moisture conditions towards the end of the summer. The seasonality of soil moisture distribution is consistent considering the seasonal changes in precipitation and evapotranspiration and has been investigated by a number of authors (Loague, 1992; Rovanešek *et al.*, 1996; Western *et al.*, 1998a; Wilson *et al.*, 2004). However, the influence of spring snowmelt on soil moisture distributions complicates the processes determining the soil moisture state with time since it is the major source of water input into the hydrological system at this time and the soil moisture patterns determined during this period persist into the early summer months.

The production of variograms (described in Section 3.2.1) has provided information on the spatial structure of the soil moisture fields observed during this study. This information, including the spatial organisation and degree of spatial autocorrelation of the soil moisture patterns, is useful when attempting to understand the processes operating across the study area. The omnidirectional variograms produced for the Hillslope site show high sills during the spring months (May and June) and progressively lower sills with time through the summer months. The Windfetch site shows a similar trend, although the jump from high to low sill values between the spring months and the late summer is much more pronounced. These findings are similar to those found by Western *et al.* (1998a) who suggest that the seasonal differences in sill values are due to the different processes controlling soil moisture in wet and dry conditions. In the wet conditions, sills are high due to the

redistribution of water due to topographic factors. Dry conditions give low sill values since soil moisture values are relatively consistent and are controlled by soil properties and evapotranspiration.

All of the variograms produced displayed a nugget effect. Modelled variograms for the Hillslope site (spherical) show nugget values ranging from 0.006 to 0.0015, the highest values being during the spring. The Windfetch site modelled variograms (exponential) showed nugget values of between 0.0055 and 0.0015. Again, the highest nugget values were observed during the spring months and decreased through the summer months. There are two main causes of the nugget effect which are measurement error and sub-grid variability. These are discussed in Western *et al.* (1998a) where small scale spatial correlations are examined using data collected at the sub-grid scale. This is a good way to determine the accuracy of the modelled variograms and to determine the cause of the nugget effect. However, this was not carried out in this study due to the limited fieldwork period but it would be recommended for future studies of this type. The nugget values observed in this study are, nevertheless, consistent with the measurement errors associated with the use of the Surface Capacitance Insertion Probe (SCIP). Western *et al.* (1998a) found that higher nugget values were observed during wetter conditions which appear to be consistent with the findings of the variogram analysis presented in this thesis.

One of the most significant results to come out of the variogram analysis is the determination of the range (correlation length) which represents the maximum distance over which spatial correlation is present. The correlation lengths determined for this study were between 50 m and 55 m. Other studies have shown considerable differences in correlation lengths. Western *et al.* (1998a) include a summary of other spatial soil moisture variability studies citing correlation lengths of anywhere between 0.7 m (Whitaker *et al.*, 1993 cited in Western *et al.*, 1998a) and 650 m (Warrick *et al.*, 1990 cited in Western *et al.*, 1998a). It is important to note that differences between correlation lengths between various studies can be due to a number of factors including the scale of the study, sampling regimes, sample sizes, measurement techniques, measurement depths and differences in topography, vegetation and climate (Western *et al.*, 1998a). However, the correlation



lengths determined for this thesis are similar to those described in the literature carried out at a similar scale. Nyberg's study of 1996, in the covered catchment at Gårdsjön, Sweden, reported a correlation length of 20 m. This was derived from 57 to 73 sample points in a 10 m x 10 m area. These samples were taken at 15 – 30 cm depth using Time Domain Reflectometry (TDR) in sandy-silty soils. Analysis presented by Petrone *et al.* (2004) show a correlation length of 10 m for a 254 m transect with a sample spacing of 0.5 m in their study of a cutover peatland in Québec. Although the examples of correlation lengths cited above are not the same as those found in Chapter 3, the differences in sampling strategy, scale of study and measurement depth could account for this. A study by Anctil *et al.* (2002) derived correlation lengths of between 31 m and 37 m for a 755 m transect at 5 m sample spacing. The soils at this site were highly organic and volumetric soil moisture was determined gravimetrically for the top 5 cm. However, the relatively small sample sizes (90 in this case) may not be sufficient to accurately characterise soil moisture patterns geostatistically. Western *et al.* (2003) suggest that to reliably quantify the correlation structure, a large number of spatial samples (around 300) should be used. However, Anctil *et al.* (2002) draw similarities between their findings and those of Western *et al.* (1998a).

Probably the best comparison of the work presented in Chapter 3 can be made with the correlation lengths described by Western *et al.* (1998a) as part of the Tarrawarra study. This comprehensive study was based on extensive field data and the thorough geostatistical analysis of these data has been widely documented (Grayson *et al.*, 1997; Western *et al.*, 1998a; 1998b; 1999a; 2003; Wilson *et al.*, 2004). Here correlation lengths were between 35 m and 60 m and were derived from 500 sample points in 10 m x 20 m area. Although there are differences in soil properties between Tarrawarra and the Abisko study area, many similarities between the variogram characteristics can be drawn. The correlation lengths determined for this study were between 50 m and 55 m which are within the range found for the Tarrawarra studies (between 35 m and 60 m). The higher nugget values observed during the spring months described in this study is also consistent with the findings of the Tarrawarra studies where higher nuggets were observed during wetter conditions. Also, the seasonality of the spatial soil moisture distributions described in this study through the histogram and variogram analysis is similar to the findings of the Tarrawarra studies.

Another aspect of the variogram analysis was the investigation of any preferred directional variability in soil moisture across the study area. The directional variograms presented in Chapter 3 (Figures 3.7 and 3.8) show that during the snowmelt period, the Hillslope site displays a higher variability in the east-west direction than in the north-south direction. The directional plots for the Hillslope site at this time are significantly different to the omnidirectional variogram, suggesting that the structure is anisotropic. However, later in the season, the directional variograms show little variation and are considered to be isotropic. Directional variograms for the Windfetch site are similar to the omnidirectional variograms at both the start and the end of the season. The directional variability displayed at the Hillslope site during the snowmelt period suggests that topography plays an important part in the soil moisture patterns observed at this time. Since the hillslope is north-facing, water is redistributed in the north-south direction and there is maximum continuity in this direction during the wet periods. Hence the directional variograms show higher variability in the east-west direction. During the drier periods, the directional variograms show no directional variability suggesting that topography is not the main driving factor for the soil moisture patterns observed. These results are again similar to those found by Western *et al.* (1998a) and Western *et al.* (2003) who found that directional variability was most pronounced when the soils at their study area were wet.

The variogram analysis indicates that there is some degree of spatial organisation to the soil moisture patterns observed at the study site during the spring. The soil moisture patterns show continuity and connectivity when the soils are wet during the spring snowmelt period. Also, the notion that two soil moisture states may be distinguished that are controlled by different processes has been suggested by Grayson *et al.* (1997) and could be relevant to the results obtained for the Abisko sites. Grayson *et al.* (1997) suggest that two preferred soil moisture states that could be determined are a wet state and a dry state which are controlled by different mechanisms. The wet state is controlled by lateral water movement in both the subsurface and surface leading to wet areas organised along drainage lines. The dry state is controlled by vertical fluxes due to soil properties and the influence of local terrain only.

Figures 3.11 and 3.12 show the soil moisture distributions derived from the spatial surveys of the Hillslope and Windfetch sites respectively. These figures show that both sites are wettest during the spring snowmelt period (May and June). At the Hillslope site, an area of elevated soil moisture was seen in the centre of the site in both years, extending downslope from areas of persistent snow cover. There is continuity of the moisture pattern in the north-south direction, which was suggested by the analysis of the directional variograms, indicating that melt water is topographically redistributed downslope. This water accumulates in a region of the grid with a shallower slope (in the north east). As the summer season progresses, the soil moisture across the grid reduces but spatial structures (drainage routes) can still be seen (especially in July and August 2004) running in the north-south direction. These structures are not as pronounced in the July 2003 survey. This may be due to differences in precipitation between 2003 and 2004. Figure 3.13 shows that July and August 2004 experienced much more rainfall than the same time the previous year. For example, in the first week of August 2004, a single rainfall event of 9 mm was recorded whereas during the same period in 2003 the maximum recorded rainfall was only 5.5 mm.

The Windfetch site shows a similar pattern of seasonal soil moisture distribution. During the snowmelt period, areas of elevated soil moisture were observed in the southern and western part of the grid. There appears to be a region running approximately east-west in the southern part of the grid where soil moisture is greatest during this time (due to the topography of the Windfetch site which will be discussed in Section 7.2). As the summer progresses, the soils dry considerably. In July, August and September 2004, the soils are wetter than the previous year (again due to higher rainfall in this year compared to 2003) but the spatial soil moisture structures are seen to persist in the southern half of the grid. The spatial patterns observed at both sites appear to support the findings of Grayson *et al.* (1997) and Western *et al.* (2003) who suggest that two soil moisture states can be distinguished (wet and dry) and that switching occurs between these two states indicating whether the process controlling soil moisture patterns is caused by lateral or vertical water fluxes. During the wet state (the snowmelt period described above), the soil is saturated and moisture patterns are controlled by the lateral movement of water (caused by topography)

which produces the anisotropy in the directional variograms and the connectivity caused by drainage routes seen on the soil moisture fields. In the dry state (July, August and September 2003), the moisture patterns are determined by local influences such as soil properties, precipitation and evapotranspiration. As the soil dries, soil water suction increases and soil hydraulic conductivity decreases. The effects of rainfall and evapotranspiration are then the main cause of the spatial soil moisture patterns. However, if there is sufficient rainfall, the soil will revert back to the wet state (as seen during the summer of 2004) and the topographic effects will cause lateral flow. This appears to be the case in this study and is most pronounced at the Windfetch site during July, August and September 2004 (when compared with 2003).

Although nine spatial surveys of soil moisture were carried out over two years, they were essentially just a snapshot of the soil moisture state on the day of the survey. In order to investigate the temporal variability of the soil moisture at the study area more fully, fine resolution, continuous time series data were collected at three locations. On the hillslope, soil moisture and temperature were measured at two locations at 5 cm and 10 cm depth; one at the base of a topographic hollow where the main land cover type was *Salix* scrub and the other at a more exposed position dominated by *Empetrum* heath. Measurements were also taken at a point approximately 100 m north east of the Windfetch site, located in a small area of bare earth in a micro-topographical depression (essentially a small 'stream' bed prone to saturation). Rainfall was also measured at the Windfetch site giving an almost continuous record for the duration of the study. The rainfall record (Figures 3.13 and 3.14) shows that during July 2004 a large amount of rainfall occurred compared to the previous year. This explains the higher soil moistures being recorded during the 2004 surveys.

The most striking aspect of the soil moisture time series data for the Hillslope site presented in Figure 3.15 is the rapid response of soil moisture to rainfall events. During the summer months, a series of wetting peaks and drying troughs can be seen at both the *Empetrum* and *Salix* sites which correspond to the rainfall events. The wetting of the soils due to rainfall is rapid and the dry-down period is typically within seven days. During 2003, these events can be seen in detail at both sites on the hillslope. At the *Empetrum* site, the 5 cm soil

moisture is considerably higher than at 10 cm depth. The *Salix* soils are wetter than the *Empetrum* soils but the difference between the *Salix* soils at 5 cm and 10 cm depth is smaller; the *Salix* soils at 5 cm depth being only slightly wetter than at 10 cm depth. The differences in soil moisture content observed at these two sites could be due to a combination of factors. The higher moisture values at the *Salix* site may be due in part to the lateral movement of water since the site is located at the base of a topographic depression. Also, differences in soil depth, soil structure and soil organic content would account for the differences (which will be discussed in section 7.3). The Windfetch time series data show, the soil at this site was saturated for most of the summer of 2004. However, given the high rainfall at this time and the location of the site (in a micro-topographical depression prone to saturation), this was to be expected.

The time scales involved in the wetting and drying cycles appear to be similar to the limits proposed by Parent *et al.* (2006). Although the soils at the Abisko site are highly organic and the data were collected on a hillslope (as opposed to the sandy/loam soils on a flat site, Parent *et al.*, 2006), the time scales of the soil drying after rainfall events (typically seven days at the Abisko sites) are within the 48 hour to 2 week time scale that accounts for the majority of soil moisture variability suggested by Parent *et al.* (2006). Parent *et al.* (2006) suggest this variability is due to the transfer of energy from precipitation to soil moisture that is dependent on the properties of the soil.

The soil temperatures for the Hillslope site (Figure 3.17) show the timing of the soil freeze and thaw cycles at this site. There are distinct differences in the soil temperature regime between the *Empetrum* and *Salix* sites during the period that the soils are frozen and during the soils thawing. When the soils are frozen, the soil temperature (at 5 cm depth) at the *Empetrum* site drops considerably below zero (down to almost -10 °C in January 2005) and is seen to fluctuate greatly. This is also the case at the Windfetch site. In contrast, the soil temperature at the *Salix* site is noticeably higher and more stable, staying at just under 0 °C during the winter months. This may be due to the greater snow depth (approximately 0.7 m deep) present at the *Salix* site during this period, since the site is situated in a topographic hollow that accumulates wind blown snow. The accumulated snow has the effect of

insulating the soil from the extremes of air temperature experienced during this time. The *Empetrum* site does not experience this effect and as a result the soil temperatures are more directly affected by the very low air temperatures. The presence and persistence of the accumulated snow in the topographic hollow of the *Salix* site during the thaw period also affects soil temperatures. Figure 3.17 shows that during the thaw of 2004, the soil temperature at the *Salix* site remained at around 0 °C whilst at the *Empetrum* site the soil temperatures rose steadily during the period. This is due to the persistence of the snow in the *Salix* site. As well as any insulating effect the snow has on the soil temperature, any melt water that enters the site due to lateral flow will cause the temperature to remain at around zero. These observations are important when considering the processes involved during the seasonal freezing and thawing of the soil and are especially relevant to the modelling of these systems (see Chapter 6, Section 6.4).

## 7.2. Topographic Influence on Spatial Soil Moisture Patterns

In order to investigate the role that topography plays in determining the spatial soil moisture patterns observed at the site, detailed topographic data for both the Hillslope and Windfetch sites were collected (see Chapter 4). The landscape of the site is dominated by the presence of solifluction terraces which cause a series of steep sided ridges and topographic hollows to be present running parallel to the contours of the slope. These ridges and hollows are interspersed by terraces of flatter topography. This gives the site a distinctly stepped appearance. Slope maps for the Hillslope and Windfetch sites derived from the topographic data (Figure 4.4) highlight these ridges, hollows and terraces. The average slope values for the Hillslope and Windfetch sites were approximately 10° and 7° respectively. At the Hillslope site, two areas of high slope can be distinguished in the southern half of the grid running in an east-west direction. Moving northward (downslope), a terrace area can then be seen, with flatter topography, before another area of high slope at the northern end of the grid. The Windfetch site is considerably flatter, with one main area of high slope occurring in the southern half of the grid running in a north-easterly direction. The presence of these localised areas of high slope are important when considering the soil moisture distributions since the topographic hollows present at the base of these slopes are



effectively sheltered from the effects of high winds and wind-blown snow. This means that larger shrubs (predominantly *Salix*) are present here and wind-blown snow has a tendency to accumulate and persist in these hollows. Also, the soil is deeper due to local deposition of leaf litter and the accumulation of wind-blown soil.

The production of Digital Elevation Models (DEMs) for the two study sites allowed the spatial soil moisture distributions (see Chapter 3) to be mapped onto three dimensional topographic surfaces. The resulting maps (Figures 4.9 and 4.10) show that topography plays an important role in soil moisture variability across the study area.

At the Hillslope site, the large wet areas observed during the spring are concentrated at the base of the central topographic 'dip' and extend downslope onto the terrace area. The wet areas persisted in the topographic hollows and also at this flatter terrace area throughout the summer. The maps for the Windfetch site show that, during the spring of 2003, the wet areas are concentrated at the base of the topographic depression running in a north-easterly direction in the centre of the grid. At the eastern edge of the grid there are two areas of low soil moisture that correspond to two raised topographic features in this area. As the soils dry out in July and August of this year, the spatial patterns of soil moisture are considerably less prominent. However, in 2004 the wet areas are again concentrated at the base of the slope but the spatial soil moisture structures persist throughout the year due to the increased rainfall experienced at this time compared with the previous year.

These findings support those presented by Kirkby and Chorley (1967, cited in Burt and Butcher, 1985) who highlight the role of topographic hollows and low-gradient slopes in the accumulation of soil water at the hillslope scale. They suggest that the hollows favour convergence of soil water flowing into the hollow and that they are a source of sub-surface runoff. The low gradient slopes favour soil saturation since sub-surface runoff is limited by the low hydraulic gradient (Kirkby and Chorley, 1967 cited in Burt and Butcher, 1985). This appears to be the case at the Hillslope site where soil water accumulates in the hollows and then redistributed downslope to the flatter terrace area.

The use of terrain indices to predict the soil moisture state due to topographic influence has been the subject of many studies (Burt and Butcher, 1985; Nyberg, 1996; Crave and Gascuel-Odoux, 1997; Western *et al.* 1999b and Western *et al.*, 2003). Here the Topographic Index developed by Beven and Kirkby (1979), derived from the DEMs of the study sites, is compared with observed soil moisture values from the spatial surveys. Although maps of the Topographic Index (Figure 4.11) suggest that saturation should occur at the base of the slopes, direct comparison of observed soil moisture values and Topographic Index values showed that the two were not well correlated. Western *et al.* (1999b) suggest that terrain indices do not represent all of the processes which can influence soil moisture patterns. Burt and Butcher (1985) note that correlation between topographic indices and soil moisture implies that other independent variables do not change across the study area. Topographic indices do not take into consideration factors that may be important in determining spatial soil moisture patterns, factors such as spatially heterogeneous soil types, soil depths and vegetation. Crave and Gascuel-Odoux (1997) highlight this and conclude that the use of only upslope and local topographic conditions is not suitable when different soil properties are present. Therefore, the lack of correlation between Topographic Index and observed soil moisture at the Abisko study sites may be due to a combination of these factors; the spatial variation in soil properties, soil depth and vegetation structure may all be contributing factors. Also, although the soil moisture patterns appear to be topographically driven during the spring season when the soils are wet, the soil moisture patterns during the drier summer months suggest that other local factors such as soil properties and evapotranspiration are important. This switch from topographically controlled lateral flow processes to local vertical flux processes has been highlighted in Section 7.1 and the analysis in Chapters 3 and 4.

The use of topographic indices has become commonplace when carrying out hydrologic investigations. Grayson and Western (2001) suggest specific conditions conducive to lateral flow need to be met in order for the assumptions on which indices are based to be valid. They argue that these conditions are rarely met and that terrain indices are often poorly correlated with spatial soil moisture variability at the hillslope scale. Hence the use of the topographic index to characterise the soil moisture state of the landscape may be useful in

specific instances where the assumptions of the index apply to the processes operating at a study site.

### 7.3. Snow, Vegetation and Soil– Effects on Spatial Soil Moisture Variability

The location of the Abisko study area (68° 17'N 18° 51'E) and its topographic features greatly influence the spatial patterns of snow, soils and vegetation structure producing a heterogeneous landscape. Although it is a low-Arctic site, the spatial distribution of snow cover in the winter months is highly variable due to the high winds which are commonly experienced. As a result, snow has a tendency to accumulate in the sheltered topographic hollows which has a profound effect on soil moisture distributions during the spring thaw period (as discussed in Section 7.2).

In order to quantify the spatial distribution of snow cover and its associated accumulation and persistence in the topographic hollows, two snow surveys were carried out over the Hillslope site in February and May 2003 (Figure 5.3). These surveys confirm that the largest snow depths occur in the topographic hollows and that, more importantly, they persist in these areas into the spring thaw. The May survey clearly shows that, whilst the other areas of the site have become snow free, the hollows remain snow covered. The soil moisture distributions described earlier are clearly driven by the input of melt water during the spring thaw and it has been concluded that water from melting snow in the hollows is redistributed due to topographical lateral flow.

The accumulation of snow in the topographic hollows across the site is also linked to the patterns of vegetation in these areas. The hollows typically contain the taller shrubs (*Salix*) since they are more sheltered and the more exposed areas of flatter topography are populated by smaller species such as *Empetrum* heath. The vegetation height survey data presented in Chapter 5 (Figure 5.6) shows that the taller vegetation does indeed occur at the base of the slopes on both the Hillslope and Windfetch sites. The taller shrubs in the hollows effectively trap the wind-blown snow causing deep snowdrifts (up to approximately 1.35 m) in these areas. This deep snowpack has the effect of insulating the

soil, causing soil temperatures to be higher than the more exposed areas (as seen in Figure 3.17). During the thaw, soils beneath the snowpack tend to remain frozen as the snow persists, causing melt water to be redistributed downslope.

The abundance of shrubs in the low-Arctic has been investigated by Sturm *et al.* (2005). In this study the concern was that the Tundra could be converted into shrubland given the warming experienced at these high latitudes in recent years. They present data that suggests that snow in areas of shrubs was consistently deeper compared to shrub-free Tundra. A positive feedback loop has been suggested by Sturm *et al.* (2005) where soil temperatures are higher in shrubland, resulting in higher microbial activity, which causes more net nitrogen mineralization during the winter and higher shrub leaf nitrogen in the summer. This may increase litter decomposition, favouring shrub growth, which traps more snow leading to higher soil temperatures (Sturm *et al.*, 2005). Hence, if taller shrubs were to become more prevalent in this region at the expense of the Tundra landscape it would have great implications on soil moisture distributions both spatially and temporally.

Also, it should be noted that a change in vegetation type would result in a change in soil moisture status, since it is clear that the soil moistures recorded at the *Salix* site were considerably higher than those at the *Empetrum* site. Cook (2005) also measured soil moisture values at areas across the Abisko site with differing vegetation types and came to the same conclusion; that soil moisture varies with differing vegetation type. There are clear links between the spatial and temporal distribution of soil moisture and vegetation. These are highlighted by Rodriguez-Iturbe (2000) who calls for a more holistic approach to hydrologic investigations that should encompass not only the hydrology of an area but also the vegetation structure, climate and soil properties to gain a better understanding of the dynamics of the system.

The soils of the study area were described by Dr. Baxter (University of Durham) and Dr. Wookey (University of Stirling) as part of the STEPPS project. Soil profile descriptions were obtained for four land cover types (*Empetrum*, *Salix*, *Sedge* and exposed ridge) at four locations. Although no direct observations exist for the Hillslope or Windfetch sites (due to

their destructive nature), these soil profile descriptions were considered representative of the land cover types sampled. The soil profile descriptions have been presented in Chapter 5 and a summary table provided in Chapter 2 (Table 2.2). All the soils, with the exception of those at the exposed ridge sites, were highly organic with high loss on ignition which decreased with depth (from 94.84 % to 35.38 %). The *Empetrum* soils showed no clear horizon differentiation and were considered to be well drained. The *Salix* soils were deeper and showed horizons O<sub>i</sub>, O<sub>e</sub>, O<sub>a</sub>, OA/OA and C. These soils were again highly organic (essentially peats), with an increase in bulk density with depth (from 0.1 to 0.27 for *Empetrum* soils and from 0.16 to 0.46 for *Salix* soils) and high loss in ignition in the top layers (94.84 % for *Empetrum* soils and 85.31 % for *Salix* soils).

Due to the heterogeneous nature of the landscape, the soil properties are considered to be highly spatially variable. Spatial measurement of soil depths (shown in Figure 5.5) across the Hillslope and Windfetch sites show that the deepest soils tend to occur at the base of the slopes and in the topographic hollows. There may be a number of reasons for this including the geomorphology of the site (the gradual creep of soils downslope due to the freezing and thawing cycles experienced at the site) and the prevalence of larger shrubs in the hollows (which produce more litter and shelter for blown soil). The soil depth will have a profound effect on the soil moisture status across the site, the deeper soils providing more soil water storage but also saturating easily and providing lateral soil water movement compared to the well draining thinner soils.

Wilson *et al.* (2004) present spatial and temporal soil moisture data and attempt to quantify sources of variability. They conclude that the spatial distribution of soils and vegetation are of similar importance to that of topography in controlling spatial patterns of soil moisture. They do note, however, that defining the spatial properties of soils and vegetation, as they relate to soil moisture, is difficult and further work should be undertaken on this problem (Wilson *et al.*, 2004).

#### 7.4. Summary & Conclusions – Soil Moisture Patterns Across a Tundra Landscape

One of the main strengths of this work has been the collection of a high quality dataset which encompasses various components of the hydrological process at the hillslope scale. This thesis builds upon the previous work in the field of spatial and temporal soil moisture variability and investigations into the processes that drive these distributions. The collection of a large spatially and temporally distributed soil moisture dataset is uncommon in this field, particularly at high latitude, low-Arctic sites. The analysis of spatial and temporal data showing the variation in near-surface soil moisture across the landscape is an essential first step to understanding the processes at work in this high latitude region. However, it is the causes of these variations that are most valued since they give an insight into the processes at work.

The statistical analysis of the spatial soil moisture variability across the site shows that the patterns are seasonal in nature. The spring period, when snowmelt and soil thaw takes place, displays higher soil moisture due to the influx of melt water into the system. It has been shown (by the variogram analysis and spatial soil moisture patterns) that this water is prone to redistribution (both surface and sub-surface) at this time. The accumulation of snow in the topographic hollows and persistence of snow in these areas, when other areas are snow free and the soils are unfrozen, is highly significant. The snow accumulation is due to the sheltered nature of the hollows and the presence of tall shrubs which act to trap the snow and reduce sublimation (Sturm *et al.*, 2005). The melt water from these distributed snow packs causes localised areas of high soil moisture downslope due to redistribution. The soil water has a tendency to accumulate in the topographic hollows and areas of low slope gradient. During the thaw period, the topography is seen to be a major factor driving the spatial variability of soil moisture at this study area.

As the season progresses, drier conditions prevail. Although spatial patterns are still distinguishable, the soil moisture variability across the site reduces and the lateral movement of soil water due to topography is not the dominant process. Instead, the local effects of varying soil properties and evapotranspiration dominate and switch the process to



one of vertical soil water fluxes. As the soil dries, changes in soil water suction and soil hydraulic conductivity become prominent and effect soil moisture patterns. If, however, there is sufficient rainfall, the soil water is again redistributed by the topography. This is seen at the Windfetch site when the soil moisture distributions for 2003 and 2004 are compared. The 2004 distributions during the summer months are wetter, due to high rainfall at this time, and the spatial variability of soil moisture increased due to lateral flows accumulating in the topographic hollows. Therefore, the evidence presented here supports the findings of Grayson *et al.* (1997) who suggest that two dominant soil moisture states can be defined (the wet state and the dry state). The processes driving soil moisture variability switch between topographical controls (when in the wet state) and local controls such as soils, vegetation and precipitation (when in the dry state).

During the annual soil freeze and thaw periods, the spatial distribution of accumulated snow has a large effect on soil temperatures. During the freeze period, soil water influx due to lateral flows in the topographic hollows acts to keep the soil temperatures around zero degrees. This is due to soil water phase changes that occur as water flows into a site with frozen soil. It has been clearly shown that, during the thaw period, accumulated snow also acts to keep the soil temperatures high (around 0 °C) and stable. This has been demonstrated by the time series data presented in Chapter 3. The *Salix* site shows much higher, more stable soil temperatures than the more exposed *Empetrum* site.

The modelling of the hydrological processes during the freeze and thaw periods using the JULES land surface scheme (Chapter 6) has been undertaken.

The determination of soil parameter values using the Monte-Carlo analysis highlighted the importance of soil properties in determining the soil moisture state at a given point.

The results presented in Chapter 6 showed that, while the  $b$  exponential derived by this method is the same as that proposed by Hall (unpublished) and Lloyd *et al.* (1999), the saturated hydraulic conductivity is an order of magnitude lower than suggested. The calibration produced 'effective' parameters which describe the water balance as though the land were homogeneous and flat and the soil parameters are uniform with depth. However,

it should be noted that, in reality, the soil hydraulic conductivity of highly organic soils is highly variable with depth and this should be considered when modelling soil moisture and temperature for these types of soils.

The main issues to arise from the modelling exercise are linked to the fact that the JULES scheme is one dimensional in nature which limits the modelling of the soil water and temperature to vertical processes only. The use of observations of soil moisture and temperature at the *Salix* site as a comparison with modelled values during the freeze and thaw periods was problematic. This was due to the fact that a one dimensional model can not predict the overlying snow conditions at a particular point in a wind-swept landscape when snow accumulation in the topographic hollows (where *Salix* is predominant) is common and deep. In the JULES scheme, the surface soil layer essentially acts as a snow/soil mix when overlying snow is present. This causes the surface soil temperature to fluctuate greatly, whereas the observed values at this time were steadier and close to zero. The influx of water into the topographic hollow of the *Salix* site due to lateral seepage would also contribute to this effect. However the model could not show this due to its one dimensional nature. When comparing the model outputs of soil temperature with those at other sites (such as the *Empetrum* and Windfetch sites), the modelled temperatures are much closer to those observed. This indicates that the variations in observed and modelled soil temperatures at the *Salix* site are due to the local effects of topography and snow accumulation in the hollows.

Although the modelling appears to have not given realistic estimates of soil moisture and soil temperature at the *Salix* site, this is due to the local effects of snow distribution and lateral seepage which are, in turn, due to the site's spatial location in a topographic hollow. When compared to other, more exposed sites, the modelling results are much more favourable. For example, when the modelled soil temperatures (Figures 6.7 and 6.9) are compared with those observed at the Windfetch logger site (Figures 3.18 and 3.19) for the freeze and thaw periods, the model appears to have performed better. This may be due to the Windfetch logger site being located on a site less prone to lateral water movement (flatter topography) and snowdrift accumulation.

The modelling of soil water during the freezing and thawing cycles is complicated due to the changing soil water suction and hydraulic conductivities during these periods of soil ice formation and depletion. Also, in a partially frozen soil, the soil moisture state is inextricably linked to the soil temperature. Therefore, modelling the soil water (both frozen and unfrozen fractions) and soil temperature in a one dimensional model does not take into account the effects of lateral seepage from adjacent areas, soil property changes with depth and spatially distributed snow cover.

This study has shown that spatial and temporal soil moisture variability across a low-Arctic Tundra hillslope is controlled by combination of processes defined by topography, spatial snow distribution, spatial soil properties, vegetation and climate. The analysis of the data collected suggests that soil moisture distributions are seasonal and vary in both time and space. In order to fully understand the processes controlling soil moisture at these high latitudes, it is necessary to measure not only hydrological but also ecological, geomorphological and meteorological processes since these factors all contribute to the soil moisture state.

### 7.5. Further Work

There are a number of recommendations that can be made in light of the findings of this thesis in order to take this work further.

#### 7.5.1. Soil Depth and Groundwater

One strength of this study has been the comprehensive field dataset; however it is acknowledged that there are many additional field measurements that could be undertaken in order to further advance this study. Although data were collected on the surface topography of the site, it would be useful to have an idea of the bedrock definition. This would provide useful topographical information which could then be used to evaluate the sub-surface accumulation and redistribution of water across the study area. This could be achieved relatively easily by surveying using ground penetrating radar technology. Also, an

indication of the groundwater levels at the site would greatly enhance this work. This was considered within this study but the installation of dipwells proved to be problematic due to the presence of a rocky/boulder layer at around 30cm depth below the soil. If dipwells could be properly installed to the correct depth at this site, an indication of the groundwater levels across the site and their temporal variability would provide useful information on the movement of water through the soils.

#### 7.5.2. Spatial Soil Properties

Detailed spatial surveys of soil properties across the study area could be carried out although this would be fieldwork intensive and difficult to achieve without destructive sampling. However, a detailed spatial description of soil properties and their variation with depth could help to explain soil moisture variability in this area. The installation of infiltrometers and tensiometers across the site would also provide additional data on soil water processes.

#### 7.5.3. Soil Moisture Survey Scales

The spatial surveys of near-surface soil moisture undertaken across the study area for this study provided data at the 10 m sampling scale for analysis by geostatistical techniques. This process could, of course, be improved by spatial sampling at much more frequent intervals. Again this would be incredibly fieldwork intensive given the size of the study sites. However, it would be useful to undertake limited sampling at nested scales (5 m, 1 m, <1 m) in order to investigate the scaling behaviour of soil moisture distributions. This information would also be useful when scaling up to the landscape, catchment and larger scales commonly used by climate models.

#### 7.5.4. Snow Melt Tracer Experiments

Since the spatial and temporal distribution of soil moisture in this region is highly dependent on the input of snowmelt water during the spring season, it would be interesting

to carry out isotope tracing experiments. These could provide data on the preferred pathways and routes of the melt water during this period and residence times of soil water in these areas.

#### 7.5.5. Terrain Indices

It was shown that the terrain index used here did not perform well. There are, however, a number of different terrain indices, some using landform curvature and others incorporating not just topographic but also other data such as soil properties and vegetation structure, that could be evaluated using the data collected as part of this study.

#### 7.5.6. Modelling Process Hydrology for High Latitudes

One aspect of this study that could be taken further is the modelling of the process hydrology at high latitudes. The evaluation of soil parameter values of different types of soil, especially organic soils, merits further study as does the variation in these soil properties with depth. However, there are two main areas that should be investigated further. The first is the ability of Land Surface Schemes to model spatially distributed snow. The effects of wind blown snow on the soil moisture distributions in this region has been highlighted by this study and, in order to produce model outputs that are representative of the processes involved, the distributions of snowpacks in the high latitudes should be included. The second challenge is that of scaling soil moisture patterns from the hillslope scale to the larger scales used by global models. This will involve some appreciation of the scaling behaviour of soil moisture fields that could be enhanced by further field studies.

In summary, this thesis has presented a comprehensive dataset of spatial and temporal soil moisture measured across a low-Arctic Tundra hillslope. The analysis of this data suggests that the spatial soil moisture variability is due to two main processes. These are the lateral redistribution of water due to topography when the soils are wet and the vertical fluxes of

water due to soil, vegetation and climate processes when the soils are drier. It is clear that the hydrology of high latitudes can have the same underlying processes as in other parts of the world. Blöschl, (2006) makes the point that the comparison of hydrological processes from case studies carried out in different regions of the world is essential in order to further the discipline of hydrology. This study has built upon those before it and has added more understanding of the hydrological processes driving soil moisture distributions in the low-Arctic. However, as always, more studies such as this should be undertaken to collect more spatial and temporal in these regions. Kane (2005) suggests that a lack of high quality data could prevent the identification of climate induced trends in the hydrology of high-latitudes. The need for continued field measurement campaigns and detailed investigation of all the components highlighted in this study is clear.



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**Appendix A**

**Surface Capacitance Insertion Probe (SCIP) – Calibration, Measurement Error and Comparison with CS616 Water Content Reflectometer**



## Surface Capacitance Insertion Probe (SCIP) – calibration, measurement error and comparison with CS616 Water Content Reflectometer

### A.1 Calibration for Soils at the Abisko Fieldsite

The Surface Capacitance Insertion Probe measures soil water content indirectly by measuring the soil dielectric constant. The two probes of the SCIP act as the plates of a capacitor with an alternating electric field. Since the capacitor is part of an oscillator circuit, when the probes are inserted into the soil the change in capacitance results in a change to the frequency of oscillation. The SCIP displays a count that is proportional to the frequency of oscillation. In order to convert the SCIP reading into the dielectric constant (also called the relative permittivity,  $\epsilon_r$ ) the following equation is used:

$$\epsilon_r = \frac{a}{R^2 - b} - c$$

(Equation A1)

where R is the probe reading (a number between 0.8 and 2) and a, b and c are specific to the SCIP being used (SCIP User Guide, 1994).

In order to obtain soil water content for a specific soil it is necessary to calibrate the SCIP gravimetrically. This was achieved by taking a reading in the field and taking a soil sample at the point of the reading. The soil corer was the same length as the SCIP rods (5 cm) giving a soil sample of known volume. The soil sample was then weighed in the lab and then dried in the oven for 24 hours (at a temperature of 105°C). The sample was then cooled in a desiccator and was reweighed. The loss of mass divided by the volume of the sample gave the volumetric water content ( $\theta$ ) of the soil. This process was then repeated for a range of soil water contents and a calibration curve for relative permittivity to volume water content was derived (Robinson and Dean, 1993).

The calculation of volumetric water content from gravimetric samples was as follows:

$$\theta = \frac{V_w}{V_t} = \frac{W_w}{W_d} \frac{BD}{D}$$

(Equation A2)

where  $\theta$  is the volumetric water content ( $\text{cm}^3 \text{ cm}^{-3}$ ),  $V_w$  is the volume of water ( $\text{cm}^3$ ),  $V_t$  is the total volume of soil ( $\text{cm}^3$ ),  $W_w$  is the weight of water (g),  $W_d$  is the weight of soil (g), BD is the bulk density of the soil ( $\text{g cm}^{-3}$ ) and D is the density of water (normally assumed equal to  $1 \text{ g cm}^{-3}$ ) (Rawls *et al*, 1993).

The default calibration capacitance probe parameters were derived experimentally from frequency readings in water and air for SCIP 06 (CEH 2004):

$$\epsilon_r = \frac{a}{R^2 - b} - c$$

(Equation A3)

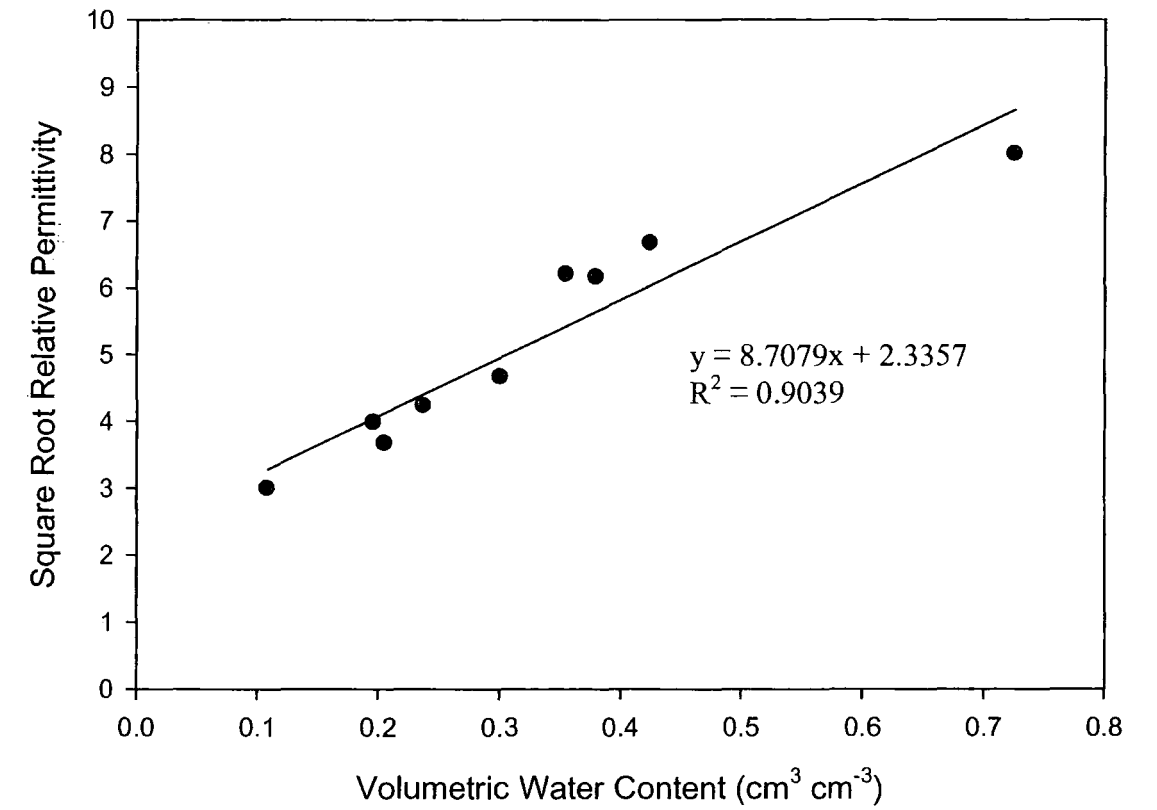
where R is the SCIP reading (in the range 0.8 – 2) and a, b and c are the probe parameter values for air and water (19.07879, 0.78645 and 7.005).

By plotting the gravimetric volumetric water content against the square root of the relative permittivity, a calibration curve was created (Figure A1).

Using a linear regression model to determine volumetric soil moisture from relative permittivity values, the calibrated volumetric soil moisture values were then found using:

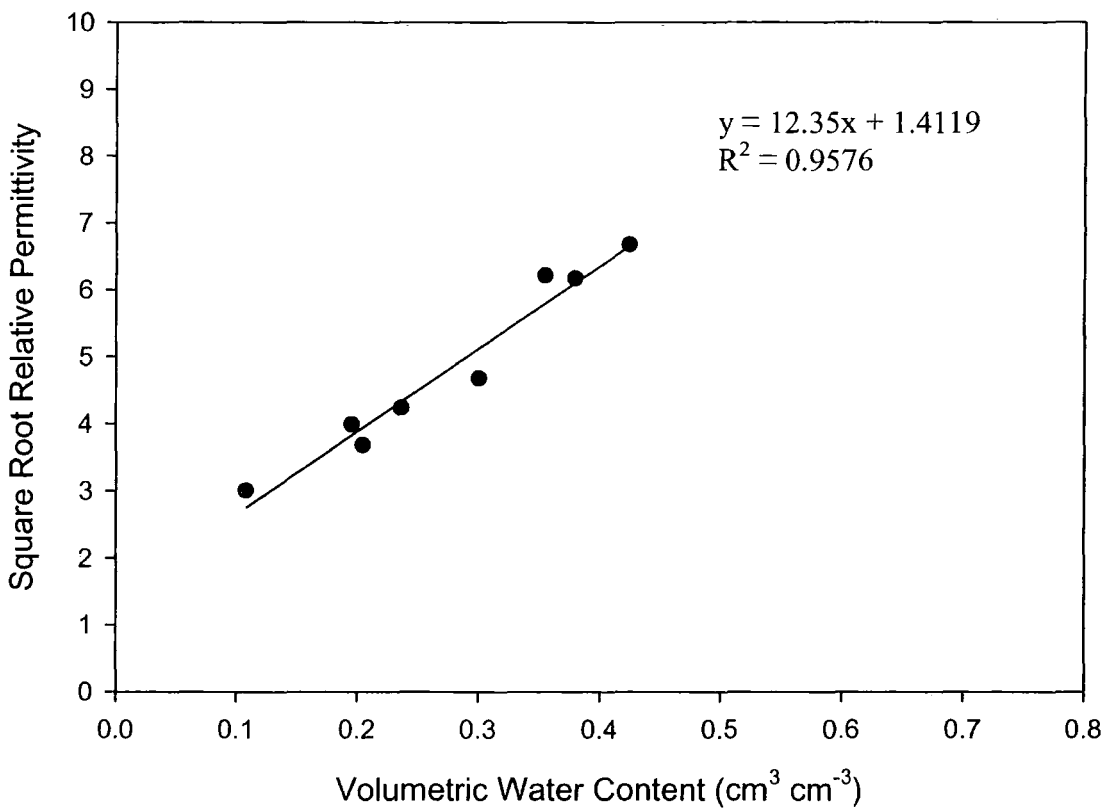
$$\sqrt{\epsilon} = a + b \theta$$

(Equation A4)  
(Gardner *et al*, 1998)



**Figure A1 Volumetric Water Content (cm³ cm⁻³) against Square Root of Relative Permittivity for Abisko Soils using a Surface Capacitance Insertion Probe**

Figure A1 shows the linear relationship between gravimetrically derived volumetric soil moisture for Abisko soils and the square root of the relative permittivity. However, the determination of the gravimetric soil moisture for saturated soils proved to be problematic. This was due to sampling difficulties of very wet soils and the transport of wet soil samples from the fieldsite to the laboratory. Therefore, it was decided that the calibration point at the wet end (0.726) should be removed from the calibration graph since the measurement errors could be large and were unquantifiable. Removing this point gave a calibration graph which can be seen in Figure A2.



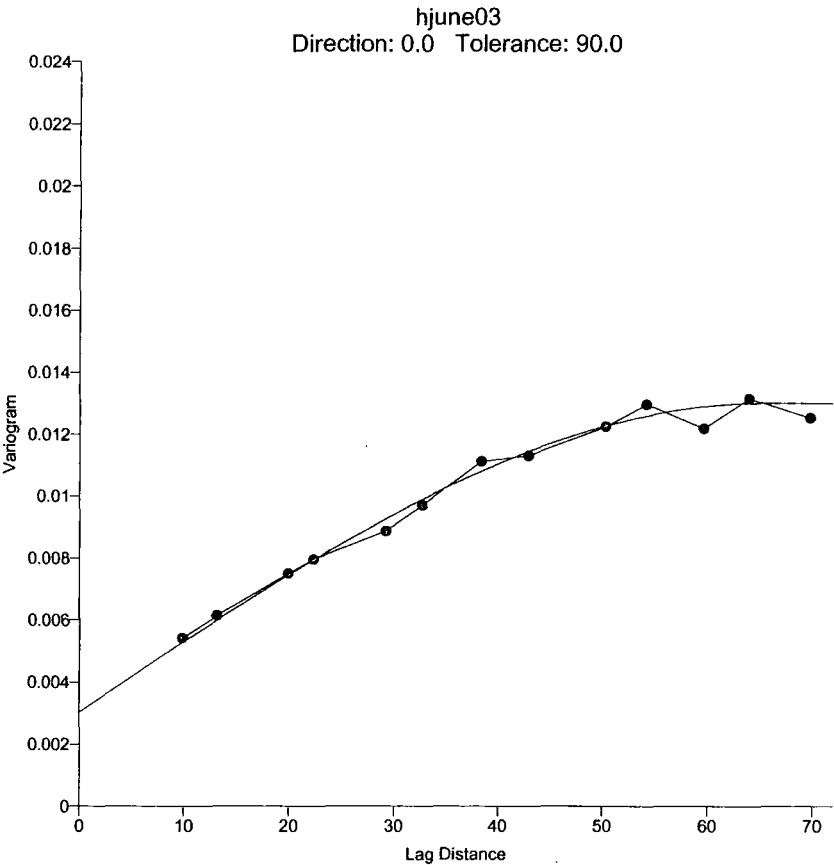
**Figure A2 Volumetric Water Content (cm³ cm⁻³) against Square Root of Relative Permittivity for Abisko Soils using a Surface Capacitance Insertion Probe (removed point at wet end)**

A.2 SCIP Measurement Error and Repeatability

In order to evaluate the measurement error and repeatability of soil moisture measurements made with the SCIP, four measurements were made at each point across the Hillslope Grid during the spatial soil moisture survey of June 2<sup>nd</sup> 2003. Each of the SCIP measurements were taken within 50cm of the grid point and the variance of the four measurements was then calculated for each point.

97% of the variance values for the entire grid (130 points – 3 excluded due to snow or frozen soil) were between 0 and 0.008 m<sup>3</sup>m<sup>-3</sup> and 90% of the variance values were between 0 and 0.005 m<sup>3</sup>m<sup>-3</sup>. This agrees with the experimental error found by Gardener *et al.* (1998) who suggest that a 0.005 volumetric water content error is unavoidable when using the SCIP.

Figure A3 shows the omnidirectional variogram (see Chapter 3) for the June 2003 survey data which includes the estimated spherical model used for the interpolation process. On examination of the variogram, the nugget observed has a value of 0.003 which is consistent with the measurement error associated with the use of the SCIP.



**Figure A3 Omnidirectional variogram for the June 2003 SCIP survey showing the estimated spherical model**

A.3 Comparison of SCIP and CS616 Water Content Reflectometer Measurements

Two sensors were used for the measurement of soil volumetric water content, the SCIP for spatial surveys and CS616 Water Content Reflectometers (Campbell Scientific, Inc) for temporal (time series) measurements. Since both sensors have been used to provide evidence of the soil moisture state across the study area throughout the thesis, it is necessary to show how the two sensors were related.

To achieve this, measurements made by the SCIP during spatial surveys on the 29<sup>th</sup> July 2003, 12<sup>th</sup> September 2003, 26<sup>th</sup> May 2004, 29<sup>th</sup> July 2004, 10<sup>th</sup> August 2004 and 4<sup>th</sup> September 2004 were compared with point measurements of soil moisture from the CS616s at the *Salix* and *Empetrum* sites recorded at the same time of the surveys on these dates. Figure A4 shows the SCIP volumetric water content against CS616 volumetric water content.

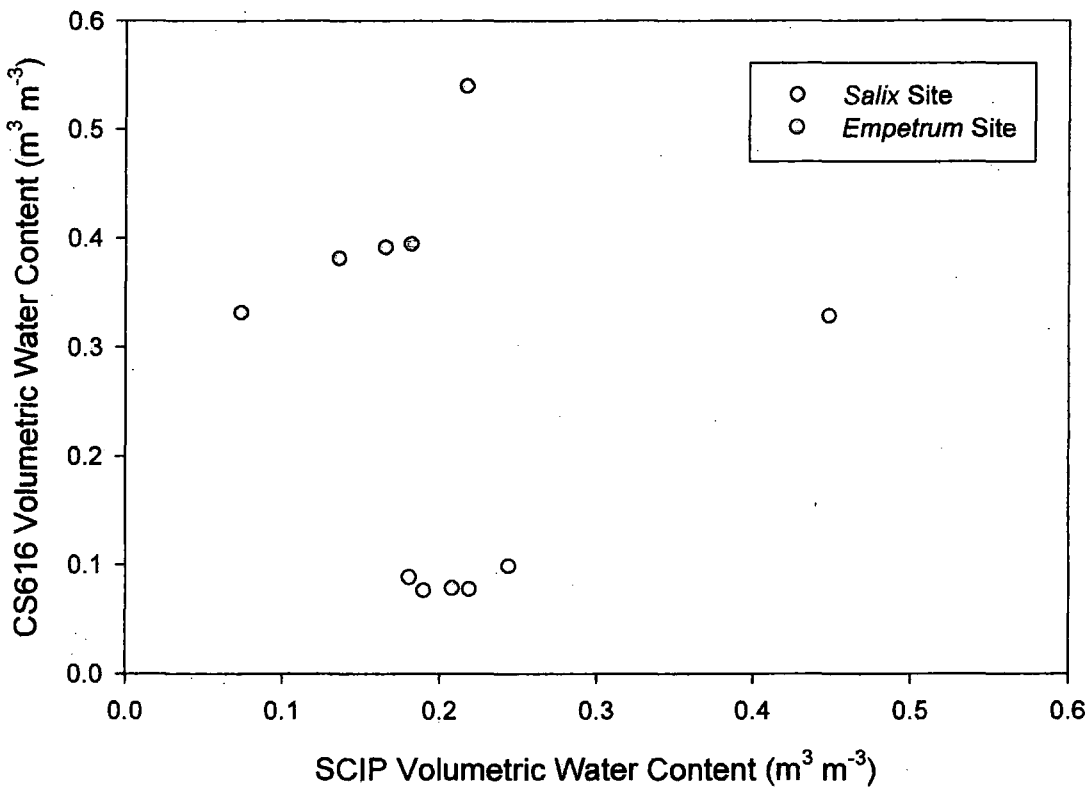


Figure A4 SCIP volumetric water content against CS616 volumetric water content on 29<sup>th</sup> July 2003, 12<sup>th</sup> September 2003, 26<sup>th</sup> May 2004, 29<sup>th</sup> July 2004, 10<sup>th</sup> August 2004 and 4<sup>th</sup> September 2004

From Figure A4, it can be seen that the CS616 probes record soil moisture values that are considerably higher than those of the SCIP. This difference can be explained by differences in both the measurement method and calibration of the two instruments. The CS616 probes measured the soil moisture every minute and then half hourly average values were recorded. It is these half hourly averages that have been used in this comparison. However, the SCIP measurements were made at a single point in time. This may explain some of the difference between the two measurements.

The main factor that can explain the difference between the SCIP and CS616 measurements is the calibration of the two instruments. The SCIP was calibrated for the highly organic soils that were present at the fieldsite (see section A1). However, the CS616s used the standard (default) calibration coefficients (for the quadratic form of the calibration equation) that were for mineral soils. Although a calibration of the CS616 sensors for the Abisko soils was attempted in the laboratory, this proved to be problematic and the results were unreliable.

Full details of the calibration equations and parameter values for the CS616 Water Content Reflectometers can be found in the user guide (Campbell Scientific, Inc, 2002).

